Institute of Polar Studies

Report No. 25

Glacial Geology of Adams Inlet, Southeastern Alaska

by

Garry D. McKenzie

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November 1970

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ABSTRACT

Adams Inlet is in the rolling and rugged Chilkat-Baranof Mountains in the eastern part of Glacier Bay National Monument, Alaska. Rapid deglaciation of the area in the first half of the twentieth century has exposed thick sections of post-Hypsithermal deposits and some of the oldest unconsolidated deposits in Glacier Bay. About 30 percent of the area is underlain by unconsolidated material; 14 percent of the area is still covered with ice. The formations present in Adams Inlet are, from the oldest to the youngest: Granite Canyon till, Forest Creek glaciomarine sediments, Van Horn Formation (lower gravel member), Adams lacustrine-till complex, Berg gravel and sand, Glacier Bay drift, and Seal River gravel. No evidence of an early post-Wisconsin ice advance, indicated by the Muir Formation in nearby Muir Inlet, is present in Adams Inlet.

Following deposition of the late Wisconsin Granite Canyon till, the Forest Creek glaciomarine sediments were laid down in water 2 to 20 m deep; they now occur as much as 30 m above present sea level. A volcanic ash unit within these sediments may have been derived from Mt. Edgecumbe. Wood at the top of this formation is dated at 10,940 ± 155 years B.P. (I-2395). Gravel infilling (lower member of the Van Horn Formation) of the valleys tributary to Adams Inlet apparently followed retreat of the sea, although there are no deposits in Adams Inlet dated between about 10,000 and 4,000 years B.P. By 3,700 B.P. these gravels had reached 3 m above present sea level in the vicinity of Adams Inlet. About 7,000 years ago gravel was being deposited near present sea level in Muir Inlet, and several episodes of lakes occurred in Muir Inlet between 4,500 and 2,200 years B.P.

Advancing ice in Glacier Bay had probably reached Reid Inlet by 4,680 ± 160 years B.P. (Y-9), and in upper Muir Inlet ice deposited till over wood about 2,120 ± 115 years B.P. (I-1610). Damming of the entrance to Adams Inlet, probably originally by outwash and later by ice from Glacier Bay, resulted in the formation of glacial Lake Adams about 1,700 B.P. The ice-dam hypothesis is supported by a date of 1,540 ± 130 years (Y-4) on wood under till on the south side of Geikie Inlet. Bottom deposits (Adams Formation) of Lake Adams were disturbed by several advances of ice from north of Adams Inlet into tributary valleys to the south. Retreat of the ice, possibly during

the Little Optimum (1150-1300 A.D.), to the central part of Adams Inlet was followed by outwash filling (Berg Formation) of the valleys south of Adams Inlet. The late Neoglacial advance formed a kame moraine at Endicott Lake where ice remained until about 1830. Several glacial lakes formed in tributary valleys during deglaciation. Lacustrine deposits of these lakes, and glacio-fluvial deposits and tills constitute the Glacier Bay Formation in Adams Inlet.

The surface elevation of the ice in Adams Inlet in 1890 was 380 m, about half of the elevation during the Neoglacial maximum. By 1941, about 15 km^2 of ice remained in Adams Inlet; this ice was less than 100 m in elevation. During this period the rate of wastage had been 7.9 m per year.

Ice-contact deposits such as eskers, kames, and pitted outwash are common; buried ice is still present in parts of Adams Inlet. Outwash deltas are rapidly building into Adams Inlet.

ACKNOWLEDGMENTS

This research was conducted as part of a continuing investigation of the glacial geology of Glacier Bay National Monument by members of the Institute of Polar Studies, The Ohio State University. The report is the result of two field seasons in the summers of 1966 and 1967 in Adams Inlet, Alaska.

The writer wishes to express his sincere thanks to Dr. Richard P. Goldthwait who suggested the problem, supervised the research, and spent several weeks in the field. Thanks are also due Dr. Arthur Mirsky, former Assistant Director of the Institute, who was helpful in many ways, including preparation for the field.

The U.S. National Park Service provided excellent logistical support during both field seasons. Special thanks are extended to Captain James Saunders and the crew of the Park Service motor vessel Nunatak for transportation to the field, and to Mr. Charles Janda and the Park Service Staff at Bartlett Cove who provided numerous resupply trips and transportation from the field at the end of the season. The enthusiastic support of the project by Mr. Robert E. Howe, Superintendent of Glacier Bay and Sitka National Monuments, and assistance from the Park Service Staff in Juneau is gratefully acknowledged.

During the 1966 field season, the writer was assisted by Mr. W. C. Lefler of the University of Kansas, and by Mr. T. R. Merrell III, of St. Olaf College. In 1967 the writer was again assisted by Mr. Merrell and by Mr. Steven Goldthwait of Dartmouth College. Assistance in establishing camp during the 1966 and 1967 field seasons was provided by Messrs. J. N. Westwater and D. I. McKenzie, respectively. My thanks go particularly to Mr. T. R. Merrell II of the U. S. Fish and Wildlife Service, who kindly helped in many friendly ways, and who provided transporation and the hospitality of his home to the field party while in Juneau. Mr. Herbert Mehrling assisted in preparation of equipment for the field; most of the drafting was done by Mr. Richard L. Jolley.

Various staff members of the Department of Geology, The Ohio State University, were consulted during the laboratory studies. Dr. Aurèle La Rocque identified the macrofossils and Dr. Walter Sweet examined some Foraminifera from the Forest Creek Formation. Dr. George E. Moore, Jr. advised on the mineralogical investigations, and Dr. Charles H. Shultz examined a sample of volcanic ash. The computer program used to reduce and plot the till fabric data was written by Dr. Charles E. Corbató.

Dr. Larry P. Wilding, of the Department of Agronomy, The Ohio State University, supervised work done by the writer in the Soil Mineralogy Laboratory. The writer is grateful for Dr. Wilding's suggestions regarding the clay minerals and the spectrographic analyses. Mr. L. Richard Drees, of the Soil Mineralogy Laboratory, assisted in processing the samples and in interpreting the results. The writer is indebted to Mr. E. Moye Rutledge, who permitted the writer to use some of the facilities of the Ohio Soil Survey Characterization Laboratory and who ran the cation analyses and several granulometric analyses. It is a pleasure to give credit to Dr. Paul A. Colinvaux, of The Ohio State University College of Biological Sciences, who placed essential equipment at the writer's disposal and who was very helpful in the palynological investigations. The writer is also thankful for advice received from Mrs. E. K. Schofield and Mrs. P. Colbaugh of the same college.

Correspondence with Dr. D. A. Brew, Dr. R. W. Lemke, and Dr. A. T. Ovenshine, geologists with the U. S. Geological Survey, greatly benefited the writer.

Members of the Institute of Polar Studies and the Department of Geology at The Ohio State University provided valuable advice and discussion. Especially helpful were: Dr. George Haselton, and Messrs. Gerald Holdsworth, John Lindsay, Cedomir Marangunić, Peter Morgan, Donald Peterson, and Geoffrey Smith.

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The manuscript was reviewed by Dr. R. P. Goldthwait, Dr. S. E. White, Dr. G. E. Moore, Jr. and Mr. J. Splettstoesser.

CONTENTS

ILLUSTRATIONS

PLATE

TABLES

INTRODUCTION

Location of Area

Adams Inlet is 108 km northwest of Juneau, Alaska, in the eastern part of Glacier Bay National Monument (Fig. 1). A detailed investigation was made of the glacial geology in the vicinity of Adams Inlet, and reconnaissance mapping using aerial photographs was carried out for all of the Juneau D-6 Quadrangle (58° 45' N - 59° 00' N, 135° 40' W - 136° 00' W). The surficial geology in upper Endicott Valley, just outside this Quadrangle, was also investigated using topographic maps and aerial photographs.

Camp Adams (58° 52' N, 136° 00' W), the base camp during the two summer field seasons, is on the south side of a peninsula in the western end of Adams Inlet (Fig. 1).

Nature of Investigations

Field work was done between June 17 and August 31, 1966, and between June 18 and August 26, 1967. The main purpose of the study was to determine the glacial history of Adams Inlet and to relate this history to that of other parts of Glacier Bay, particularly the Muir Inlet region. To carry out this task a detailed map (Plate I) of the surficial geology was made, and stratigraphic sections in unconsolidated material were measured.

Mapping was done on a Juneau D-6 map sheet (1:63,360) and on aerial photographs on a scale of approximately 1:40,000. Field work was concentrated in the valleys where most of the unconsolidated deposits occur. A thickness of 0.5 m was taken as the boundary between glacial drift, mostly till, and bedrock with a thin cover of drift, except in the cases of ground moraine or lacustrine deposits over unconsolidated material. These surficial covers are shown even where they are as little as 0.1 m thick. In inaccessible areas, where the nature of the unconsolidated deposits is uncertain, the deposits are marked "U."

Where possible, changes in rivers, lakes, and shorelines, compiled from recent unpublished aerial photographs by A. Post, U.S. Geological Survey and from the writer's field observations, have been included in the maps accompanying this report. Changes in glacier margins have been restricted to Casement and Adams Glaciers, which show the greatest differences from the 1949 map (compiled from aerial photographs taken in July, 1948).

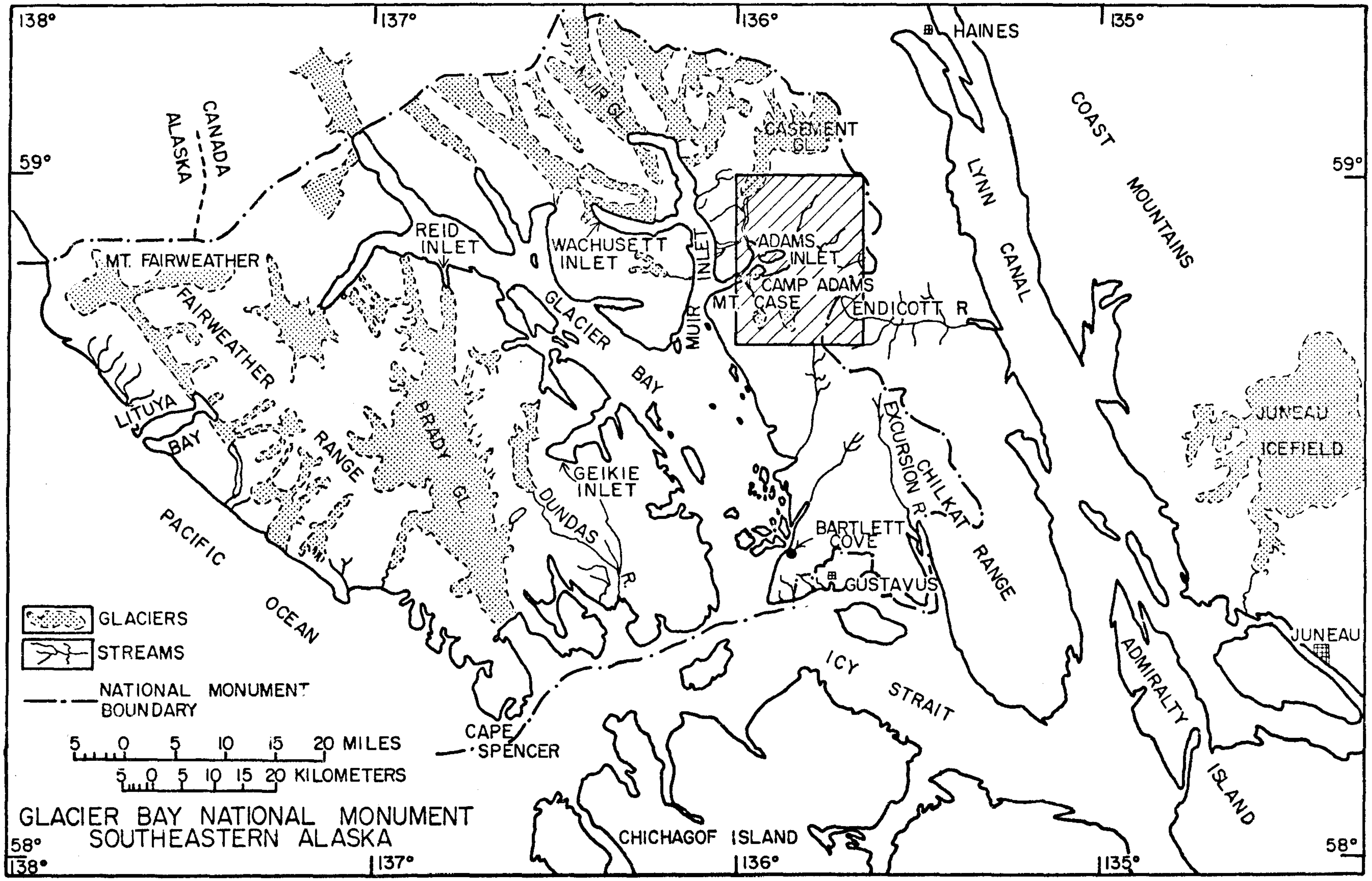


Figure 1. Index map of Glacier Bay National Monument. Diagonally-lined part is the area of Figure 2.

Ninety-seven stratigraphic sections (Fig. 2), varying in thickness from 5 to 250 m, were studied. Most of the sections were sampled; samples included bulk till, pebbles from till and gravel, sands, clays, wood, and shells. Only the key sections have been included in this report; however, a description of all stratigraphic sections is on file at the Institute of Polar Studies, 125 South Oval Drive, Columbus, Ohio 43210.

In an attempt to differentiate till units and to determine source areas, several properties were analyzed: pebble composition, till fabrics, grain size of the < 2 mm fraction, carbonate content of the < 200 mesh grade, clay minerals, and composition (as determined by X-ray spectrography) of the sand and silt fractions. Details of those methods are given in McKenzie (1968).

Previous Investigations

Many observers have recorded the retreat of the glaciers and have worked on the glacial history of Glacier Bay, but relatively few have worked in Adams Inlet. A brief account of exploration and scientific work in Glacier Bay with particular reference to Adams Inlet is given here; a more detailed summary of observations is given by Field (1947).

The first observations in Glacier Bay were made in 1794 by Mr. Whidbey, a member of Captain Vancouver's party. Vancouver (1801, and quoted in Wright, 1891, p. 55-57) recorded that Glacier Bay was completely covered with ice to a point somewhere north of what is now Bartlett Cove (Fig. 1).

John Muir was one of the first to explore the upper part of Glacier Bay, particularly that arm of Muir Inlet which is now Adams Inlet. Muir (1915) visited the area in 1879, 1880, and again in 1890 with H. F. Reid. In 1880 Muir made a sledge trip on the eastern part of Muir Glacier (Adams Inlet Glacier), and climbed Tree Mountain noting the tree line of mountain hemlock. Besides his ecological observations he also described glaciers, lakes, moraines, and ice-flow directions. He suspected a divide between Endicott Valley and Muir Inlet, and called this part of the glacier "Divide Glacier." This position is now the low area of Adams Inlet; the divide between these two regions is Endicott Gap.

G. F. Wright (1887) visited Glacier Bay in 1886, and produced a sketch map of Muir Glacier showing the terminus and moraines. He also noted the flow direction indicated by light-colored bands in the ice, and barometrically determined the elevation of the ice to be 320 m (1,050 ft) over the present position of Adams island.* Adams Inlet

*Geographic features with unofficial names are not capitalized in this report.

Figure 2. Index map of stratigraphic sections.
Contour interval in feet

Glacier, moving more slowly than the main part of Muir Glacier, was much smoother, which is probably the reason that most of the earlier visitors made trips and observations on this part of the glacier.

The early visits of H. F. Reid (1892, 1896) were in 1890 and 1892 when he constructed the first detailed map of the area (Fig. 3), on a scale of 1:150,000. In addition to carefully mapping most of the area of Adams Inlet, he described the former ice positions, the date of the last advance and the motion of the glaciers. His excellent photographs are particularly valuable in reconstructing the history of deglaciation. He showed remarkable foresight, and among his predicted changes was the deglaciation of Adams Inlet by 1946. According to Field (1964), the last portion of glacier covering Adams Inlet disappeared between 1941 and 1950. H. P. Cushing (1891), who accompanied Reid in 1890, also made valuable observations on the tributaries of Muir Glacier, recent recessions, and glacial deposits.

Maps made by the Canadian Boundary Survey in 1895 (Alaskan Boundary Tribunal, 1904) are also useful in reconstructing glacier positions, as are the first aerial photographs of the area made by the U.S. Navy Alaskan Survey Expedition in 1929.

W. S. Cooper (1937) made several visits to Glacier Bay, his first being in 1923. He described the stratigraphy and made observations on earlier positions of the ice, trim lines, and the rates of ice retreat, particularly in Muir Inlet. He also made valuable observations on forest successions, and the study of plots established by him has been continued by Lawrence (1958).

W. O. Field, who first visited the area in 1926, recorded the ice retreat in Glacier Bay by means of photographs at survey stations. These photographs, along with others on file at the American Geographical Society, are the most important information on the ice retreat in Glacier Bay. Observations and photographs of the deglaciation of Adams Inlet from 1880 to 1946 are recorded in a report by Field (1947). More observations on ice positions were made by Field (1964) in Adams Inlet in 1950 and in 1958 with R. P. Goldthwait, who noted the stratigraphy of the Inlet.

D. J. Miller of the U.S. Geological Survey described several sections on the south side of Adams Inlet in 1958. Three of the samples collected by Miller were later submitted for dating by R. P. Goldthwait. Recent work by the U.S. Geological Survey in Glacier Bay included observations on the glacial geology by A. T. Ovenshine (1967) during a bedrock mapping program headed by D. A. Brew.

Much of the recent work in Muir Inlet has been carried out by members of the Institute of Polar Studies. Goldthwait (1963, 1966, Goldthwait and others, 1966) worked on the glacial stratigraphy of Muir Inlet and provided a base for the glacial chronology of Glacier Bay

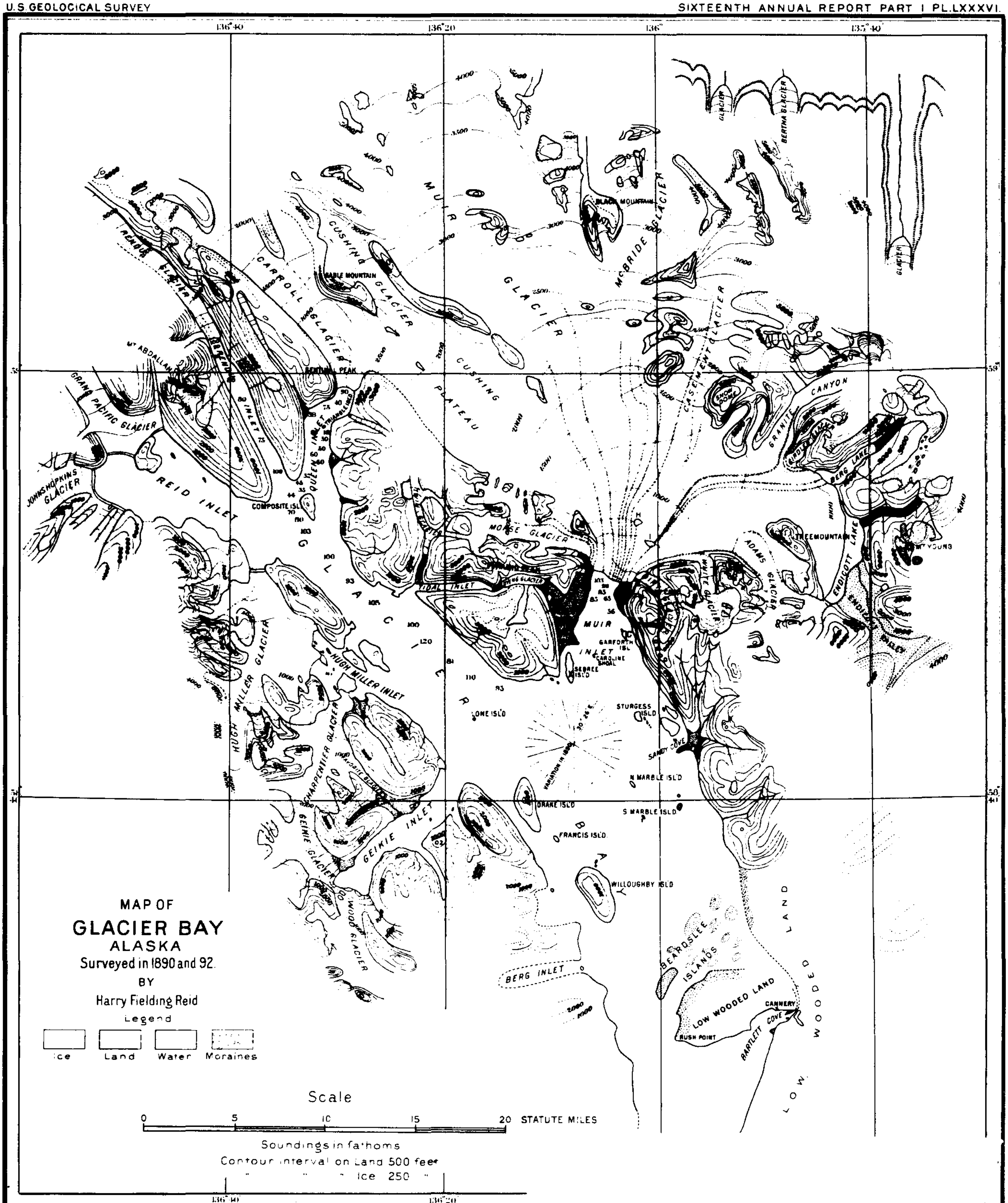


Figure 3. Map of Glacier Bay in 1890-1892 by H. F. Reid (1896)

using radiocarbon dating of tree stumps. In 1965, as part of an ecology survey (Goldthwait and others, 1966) Goldthwait sampled and described sections in Adams Inlet and near Casement Glacier, and took aerial photographs of Adams Inlet. L. D. Taylor (1962) studied ice structures on Burroughs Glacier west of Muir Inlet in 1959 and 1960, and R. J. Price (1964, 1966) described the formation and morphology of eskers at the terminus of Casement Glacier to the north. The glaciology of Casement Glacier was studied by D. N. Peterson (1968, 1969) during the 1965, 1966 and 1967 field seasons. Most formation names for glacial deposits in Muir Inlet were proposed by G. M. Haselton (1966) during a detailed study of the glacial geology of upper Muir Inlet.

Glacial Geology

In Adams Inlet Wisconsin and younger deposits only are represented in the stratigraphic column, with the bulk of the deposits being Hypsithermal and Neoglacial in age. The oldest glacial deposits yet discovered in Glacier Bay occur in Adams Inlet; they are the Granite Canyon and Forest Creek Formations, which are older than 10,940 ± 155 years B.P. in Adams Inlet. Dated deposits between 10,000 and 4,000 years B.P. are apparently absent in the area.

The remainder of the deposits in Adams Inlet, from oldest to youngest, are (a) the Van Horn Formation, lower gravel member, (b) Adams lacustrine-till complex, younger than 1,700 ± 100 years B.P., (c) Berg sand and gravel, and (d) Glacier Bay drift. Outwash deposits still forming today by retreating glaciers are known as the Seal River Formation; alluvium and colluvium constitute post-glacial deposits of the area.

DESCRIPTION OF AREA

Physiography

Glacier Bay National Monument is a mountainous area cut by many deep fiords, some of which contain tidewater glaciers. The Monument, with an area of approximately 9,000 km^2, is as much as 140 km long and over 110 km wide. The main body of water is the northwest-trending Glacier Bay (100 km $\times$ 20 km), which is over 300 m deep. Approximately 14 percent of the area of the Monument is made up of Glacier Bay and its arms and tributaries; 21 percent of the area is covered by glaciers.

According to Wahrhaftig (1965), who has described the physiographic divisions of Alaska, southeastern Alaska lies within the Pacific Mountain System, which includes the Coast Mountains, Coastal Trough, and Pacific Border Ranges (Fig. 4). All of Glacier Bay National Monument is within the Pacific Border Ranges province, which includes the highest peaks in North America, some with elevations of 6,100 m and many with elevations between 2,400 m and 3,600 m. This province is divided into the Gulf of Alaska coastal section, the St. Elias Mountains represented in the Monument by the Fairweather Range, and the Chilkat-Baranof Mountains.

Most of the Monument is a highland of varied topography known as the Chilkat-Baranof Mountains (Section III on Fig. 4). They are divided into 3 subsections. Adams Inlet is surrounded by the rugged Alsek Ranges subsection with elevations from 1,200 m to 2,200 m. The Glacier Bay subsection, an area of low rounded mountains with drowned topography, varies from 300 m to 1,500 m in elevation. A portion of this subsection in the vicinity of Bartlett Cove is an outwash and ground moraine lowland less than 300 m in elevation. A third subsection, comprising the southwest part of the Monument, is the Chichagof Highland consisting of accordant rounded summits 900 m to 1,000 m in elevation, and long fiords and through valleys.

The area of study, the Juneau D-6 quadrangle (Fig. 2), comprises 536 km^2 of which Adams Inlet occupies 5 percent. Mountains and glaciers cover most of the quadrangle except for the lowlands adjacent to the Inlet and in the tributary valleys. Over 14 percent of the area was covered by glaciers in 1967, the largest of which is the Casement in the northwest corner of the quadrangle. Unconsolidated deposits, which occur mainly in the lowlands, comprise 30 percent of the area. Lakes are small in number and their areas total 0.6 km^2. The two largest ones are Salmon lake on Adams island, and Endicott Lake. Most large rivers and streams originate at glaciers. The longest river is in Granite Canyon; the widest and most treacherous outwash rivers are the Seal, Adams, and Goddess.

Rapid retreat of ice from part of the Monument is being followed by reforestation. Vegetation present at any one part of the Monument is

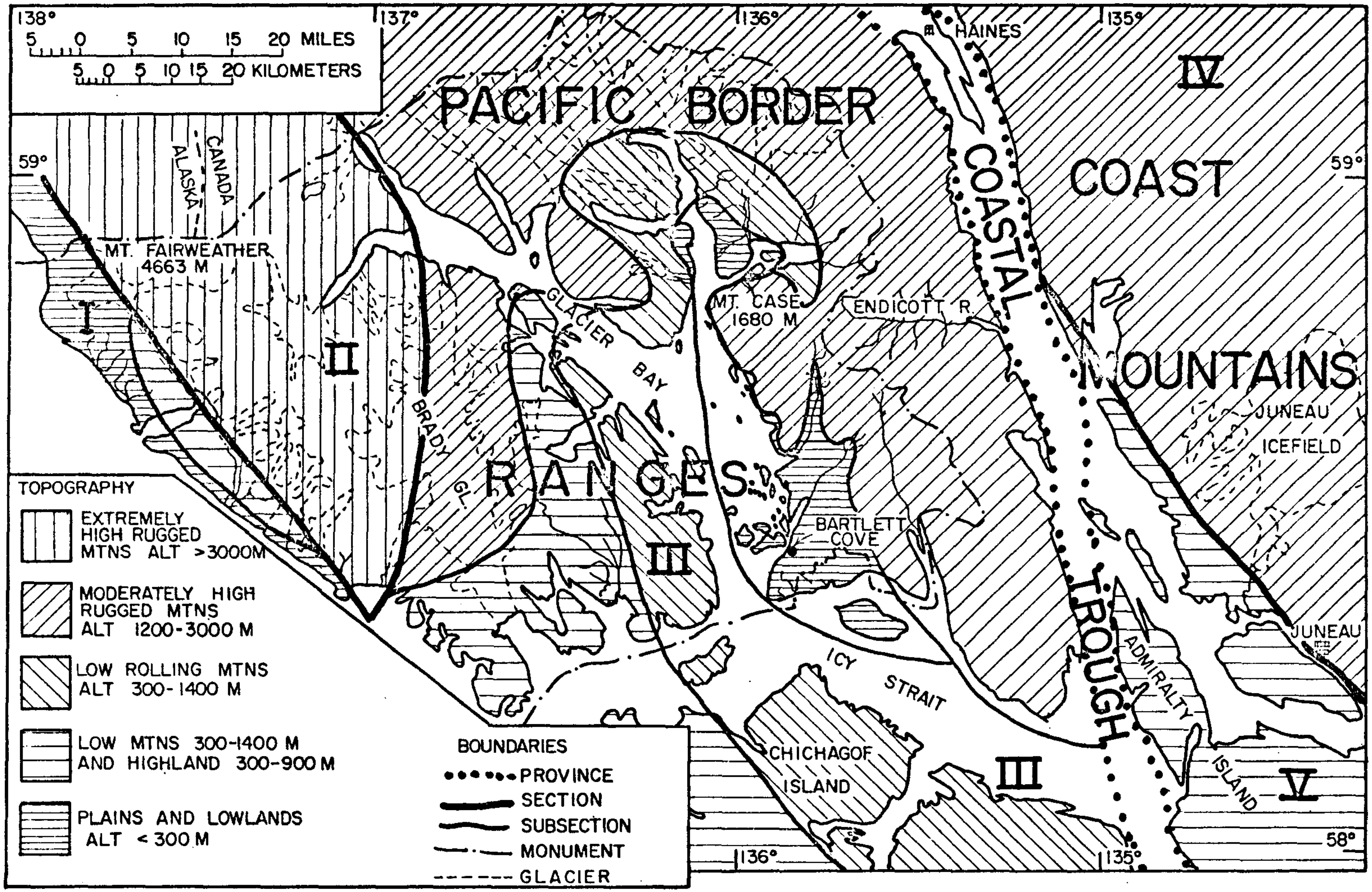


Figure 4. Physiographic divisions of part of southeastern Alaska. All of southeastern Alaska lies within the Pacific Mountain System; Glacier Bay National Monument is confined to the Pacific Border Ranges province. Roman numerals indicate sections within the provinces: I - Gulf of Alaska coastal section, II - Fairweather Range, III - Chilkat-Baranof Mountains, IV - Boundary Ranges, V - Coastal Foothills. Subsections indicate general topography. (Modified after Wahrhaftig, 1965)

thus a function of the length of time since deglaciation. Based on a study in Muir Inlet, H. F. Decker (in Goldthwait and others, 1966) has set up 9 intergrading stages of vegetation ranging from Early Pioneer stage with Dryas and Salix seedlings near the Casement Glacier terminus, to the Climax stage of spruce-hemlock forest at Bartlett Cove. In the area of study (Plate 1) all stages are represented except the Climax stage, which probably occurs just south of the map in Beartrack Valley. The Endicott Gap region (Fig. 2) is in the Closed Thicket and Poplar Line stages, making the area almost impenetrable. Much of the lowland area in Adams Inlet is in the Open Thicket stage, although the Poplar Line stage is present in some places.

Wildlife in the Muir Inlet area has been described by Welch (1965), and by several members of the Institute of Polar Studies (Goldthwait and others, 1966). Wildlife observed during the 1966 and 1967 field seasons in the Adams Inlet area was recorded by Merrell (in preparation).

Bedrock Geology

Bedrock in the vicinity of Adams Inlet consists of Paleozoic clastic and carbonate sediments, metasediments, volcanics, metavolcanics, and diverse intrusions (Fig. 5). The most recent information available is in a U.S. Geological Survey open-file report on the geology of Glacier Bay National Monument (MacKevett and others, 1967). Little work had been done previously on the geology of Adams Inlet, although Muir Inlet to the west has been studied by Cushing (1891), F. E. Wright and C. W. Wright (1937) in 1906 and 1931; Reed (1938); Twenhofel (1946), who worked on the east side of Muir Inlet, and Rossman (1963), who mapped much of Glacier Bay, including the lower part of Muir Inlet. Haselton (1966, 1967) investigated outcrops in the vicinity of Muir Inlet and summarized earlier reports on the geology of that area.

The lithologies in the area of Adams Inlet have been grouped according to their position south, east, and north of the Inlet by Brew (written communication), from whom most of the following descriptions are taken. The rocks south of Adams Inlet between Muir Inlet and Endicott Gap are composed of unmetamorphosed detrital clastics, carbonates, and volcanics of Early Silurian through Middle Devonian age. There are also many diabase dikes from a few centimeters to a few meters thick. The dominant clastic rocks are thick-bedded to rhythmically thin-bedded calcareous graywackes. At Mt. Wright the graywackes are interbedded with gray limestones; near Tree Mountain the clastic rocks are thin-bedded siltstone and shale. Outcrops of several types of unfossiliferous limestones and grayish-green amygdaloidal basalt occur near Mt. Case and Mt. Wright.

At the east end of Adams Inlet between Endicott Gap and Granite Canyon metavolcanic rocks are dominant with some slightly metamorphosed detrital clastics and carbonates. These rocks are probably the same

EXPLANATION
(for Figure 5)

Surficial deposits

Contact

Detrital clastic rocks
(some metamorphosed)

Fault

Carbonate rocks
(some metamorphosed)

Volcanic and metavolcanic rocks

Heterogeneous gneisses

Foliated granitic rocks

Unfoliated granitic rocks

Gabbroic rocks

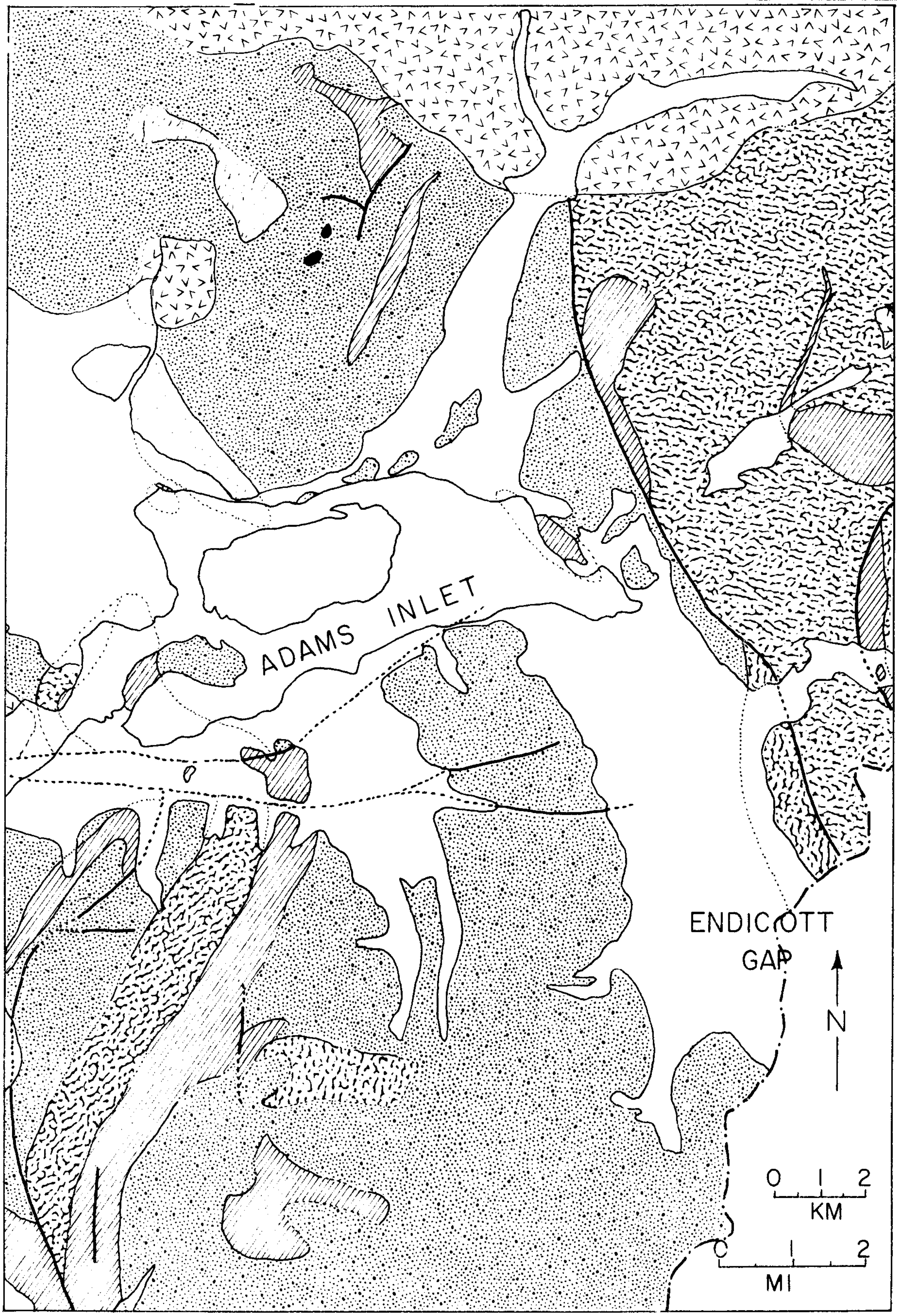


Figure 5. Bedrock geology. (Modified after MacKevett and others, 1967)

age as those south of Adams Inlet; the metavolcanics are cut by granite intrusions. The volcanic rocks of green, gray, and reddish-gray greenschists, some of which are recognizable flows, are interbedded with limestone and graywacke units.

The limestones are locally marbleized, and at the east end of Adams Inlet they contain graptolite-bearing shale partings. At Berg Creek reddish coarse-grained poorly-foliated granites occur. Dense mafic dikes and granodiorite are also present along Granite Creek.

In the area between Granite Canyon and Muir Inlet granodiorite and hornfels are dominant with some minor intrusions, carbonates, and volcanics. The granodiorite is heterogeneous with abundant porphyritic phases, and altered in some places. Syenite, gabbro, and hornblende gabbro intrusions occur north of Adams Inlet, and a quartz monzonite body is located near the head of Casement Glacier. Calc-silicate hornfels and marbleized limestone are common. There are a few areas of unmetamorphosed clastics, and southwest of Red Mountain there is a body of gray medium-bedded limestone with fossils. Amygdaloidal flows similar to those south of Adams Inlet crop out on the north shore of the Inlet, and a greenstone body occurs in Forest Creek.

North of Casement Glacier and Muir Inlet are unmetamorphosed volcanics, detrital clastics, and thin carbonates. Along the Monument boundary lies a large mass of well-foliated quartz diorite.

Climate

Glacier Bay National Monument lies in that portion of southeastern Alaska dominated by maritime influences, and thus characteristic features of the climate are small temperature variations, high humidities, high fog frequencies, much cloudiness, and abundant precipitation. The rugged topography and narrow inlets result in much local variation in temperature and humidity.

According to Köppen's classification of climates, most of southeastern Alaska near sea level belongs to the 'Cfc' type. This is a warm temperate rainy climate characterized by the absence of a dry period and having a cool summer (1 to 4 months with mean temperature greater than 10°C). The temperatures for the coolest months at Juneau (-2.3°C), Haines (-4.8°C), and Gustavus (-3.4°C) indicate that for the same time most of Glacier Bay has a temperature less than -3°C. This would place Glacier Bay within the 'Dfc' (Cool snow-forest climate, moist in all seasons with cool summers) climatic type of Köppen.

May and June usually have the most pleasant weather as is shown by the occurrence at Juneau of the greatest amount of sunshine in May and the smallest monthly precipitations in May and June. Maximum precipitation is during the months of September, October and November. The average total precipitation for this period at Cape Spencer is 115 cm, for Juneau it is 78 cm, and for Gustavus in between it is 57 cm.

Weather Observations in Adams Inlet

Weather observations in the Muir Inlet area have been made by expeditions since 1959, although the periods of observation and the elements observed have not been consistent. The data for these observations have been reduced by Loewe (in Goldthwait and others, 1966). A summary of the weather observations for Adams Inlet is given below; details are given in McKenzie (1968).

Results of observations from the summers of 1966 and 1967 in Adams Inlet suggest that it is warmer than Muir Inlet and other localities to the north. The July average for the two seasons in Adams Inlet is 12.2°C, 2.4°C above the July average for the other stations near sea level in this part of the Monument. The absence of nearby glaciers, the presence of trees and shrubs, and the generally sheltered nature of the Inlet probably are the causes of the higher temperatures. For the two years of observations the difference in mean maxima between Juneau and Adams Inlet for July is approximately 1°, and for August it is approximately 1.8°C, with the lower temperature occurring in Adams Inlet. The mean minima for July at Adams Inlet is less than 1° lower, and for August approximately 1.5°C lower than at Juneau. The January mean in Adams Inlet is probably within the limits of the means for Gustavus (-3.4°C) and Haines (-4.8°C) because it is midway between these two stations and at sea level.

The average precipitation for July 1966 and 1967 is 0.24 cm per day. The average for the latter part of June is lower, and for August it is higher. In August 1966 there were 25 days with more than 0.025 cm of precipitation; the maximum precipitation in 24 hours during that month was 5.49 cm.

The mean relative humidities for Adams Inlet in July 1966 and 1967 are 87 and 89 percent, respectively. In Juneau for the same months the relative humidities were 83 and 89 percent, respectively. The lowest relative humidity observed was 24 percent on August 26, 1966.

The dominant wind directions during the times of observations were from the west and southwest. Easterly winds were most common in the mornings. Winds were generally weak (less than 0.8 m per sec) during observation times, but were stronger during the day. Wind speeds during several periods of gustiness were estimated at 7 - 9 m per sec.

The average cloudiness (in tenths) for the two summers, from estimates at 0700 and 2100 hours, is 8.7 for July and 8.5 for August. During the summer of 1966, 72 percent of the days were mostly overcast, whereas only 10 percent of the days were mostly clear.

Hours of sunshine were visually estimated during the summer of 1967. July had the highest daily average with 4.8 hours out of a possible 15.5 hours. There were 8 days without sun during this month.

Composite Section

Deposits of till, marine clay, outwash gravel, and lacustrine clay with associated till occur in Adams Inlet. Outwash gravels and lacustrine clays are most readily visible and make up the greatest volume of unconsolidated deposits. A thin blanket of till covers most of the unconsolidated deposits and some bedrock.

No published data are available on the deposits of Adams Inlet although sections from the south side of the Inlet have been described by Goldthwait[1] and Miller[2]. Haselton (1966, 1967) has described the glacial geology of nearby Muir Inlet with which the deposits in Adams Inlet can be compared. Six units are present in Muir Inlet. In ascending order they are the Forest Creek Formation, Muir Formation, the lower, middle and upper units of the Van Horn Formation, and the Glacier Bay till. Of these units only the Forest Creek, the lower unit of the Van Horn, and the Glacier Bay Formation are present in Adams Inlet. The Muir Formation, which rests on the Forest Creek Formation at Forest Creek, and on bedrock at several other localities in Muir Inlet, was not found in Adams Inlet. Other units present in Adams Inlet are Granite Canyon till, Adams lacustrine silt and clay with a till member, Berg sands and gravels, and deposits of glacial Lake Endicott. Recent outwash deposits have been described as the Seal River Formation (Goldthwait and others, 1966). A composite section of the surficial deposits in Adams Inlet is given in Figure 6.

Granite Canyon Till

The name Granite Canyon Formation is proposed for the lowest till exposed in Granite Canyon. The till is a dense dark gray loam with abundant pebbles and cobbles. The type locality (58°55'N, 135°50'4"W) is 0.8 km northeast of the mouth of Granite River as shown on the 1949 Juneau D-6 map or 1.1 km northeast of the present mouth of the river. Here the till rests on striated and grooved limestone, and grades upward into the Forest Creek glaciomarine clay and silt (Fig. 7). The till, 2 m thick, is exposed for 30 m along the northwest side of Granite River and for approximately 20 m along Goddess river at section 42. No other exposures in Adams Inlet are known.

[1]Field notes, 1965, The Ohio State University
[2]Field notes, 1958, U.S. Geological Survey

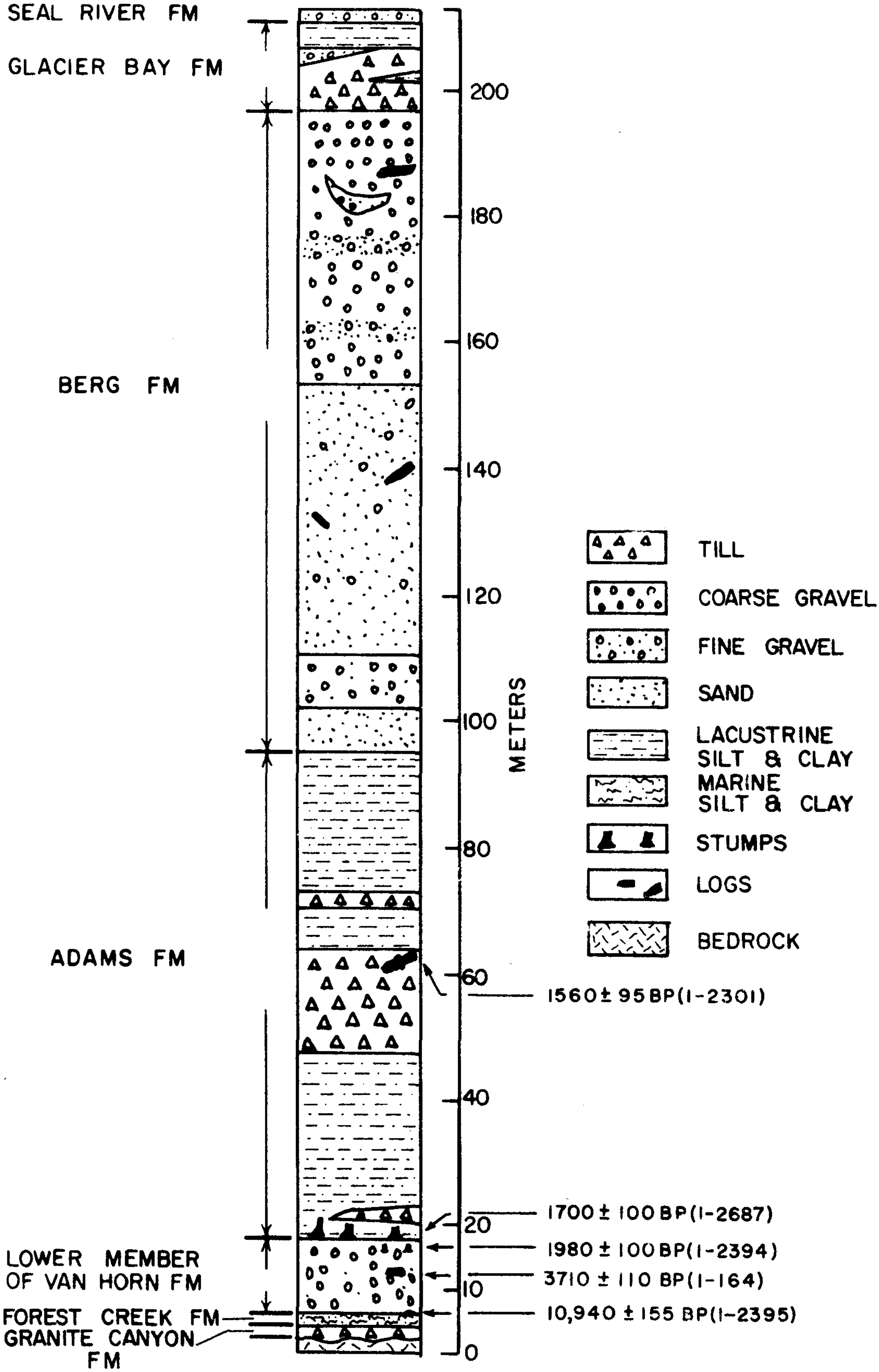


Figure 6. Composite stratigraphic section for Adams Inlet area.

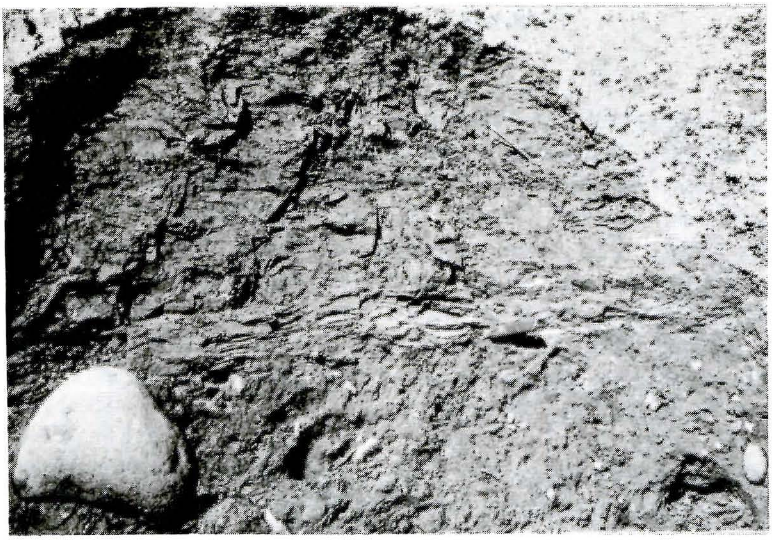
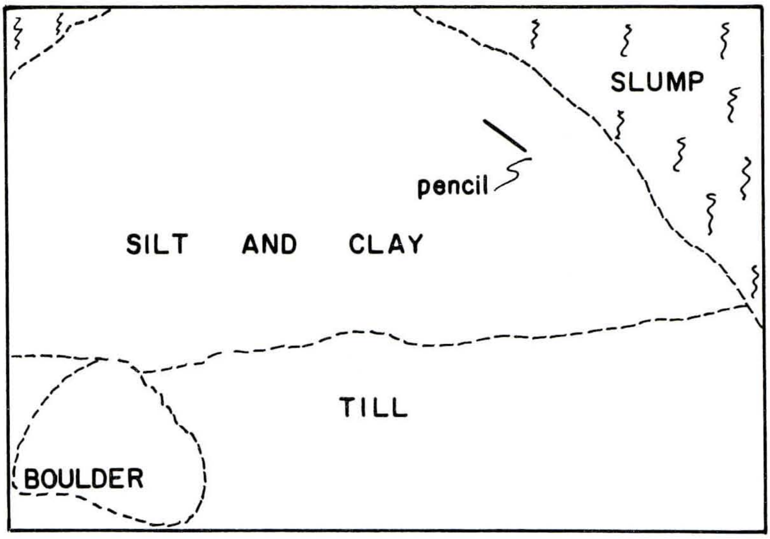


Figure 7. Contact, indicated by dashed line below, between Granite Canyon till and Forest Creek silt and clay at Section 67A in Granite Canyon. Note pebbles in till and shells (near pencil) in marine clay and silt. Materials are indicated by the line drawing below.



Mechanical analyses on four samples of Granite Canyon till, two from Granite Canyon, one from near Endicott Gap, and the third from a till discovered by the writer at the base of Haselton's Section 7 (Haselton, 1966) on Forest Creek cluster together on a triangular coordinate diagram (Fig. 8). The average of these four samples is: 41.0 percent sand, 39.2 percent silt, and 19.8 percent clay (Table 1). The till is classified as a loam or clay loam, and on the basis of the few mechanical analyses run, it apparently can be differentiated from the other tills in the area by mechanical analysis (Fig. 8). (See also the section on the Glacier Bay Formation.)

Pebble counts were made on four samples of Granite Canyon till from Granite Canyon and on two from near Endicott Gap. Although there are wide variations in the individual lithologies from both areas the till is characterized by low plutonic igneous and low limestone pebble percentages, and a high percentage of graywacke (Table 2).

The lack of plutonic igneous pebbles in the Granite Canyon till reflects derivation of this size fraction from local sources (Fig. 5). In an attempt to determine differences between the pebble lithologies of the tills and gravels of the map area, the ratio of percentages of pebbles of plutonic igneous rocks to pebbles of metasedimentary and sedimentary rocks (including limestone, graywacke, and hornfels) was calculated for each sample from a formation. These PI/M ratios reflect the number of pebbles derived from north of Adams Inlet where the plutonic igneous rocks crop out (Fig. 5). For the six samples of Granite Canyon till the PI/M ratio is 0.02 and is the lowest ratio of the three tills in Adams Inlet. Little difference exists in the relative abundances of pebbles in the two valleys (Table 2).

The relative abundance of pebble types in the Granite Canyon till in Adams Inlet is much different from that of the Muir till (Wisconsin glaciation) in Muir Inlet (Haselton, 1967, Appendix I). More plutonic igneous rocks, reflecting the lithology of the source area, occur in the pebble counts of Muir till. In four samples of Muir till, omitting two samples from upper Forest Creek (discussed later), that may not be till, the average PI/M ratio calculated from Haselton's data is 0.54. These different ratios, 0.54 and 0.02, from the Wisconsin tills in this part of Glacier Bay National Monument suggest two separate ice lobes having different source areas. Although relative abundance of pebble types will change with distance from the source, because of disintegration and incorporation of other lithologies during transport, the differences between pebble counts from Muir and Adams Inlets appear too great to indicate that the ice masses that deposited the Muir and Granite Canyon tills were of the same flow system.

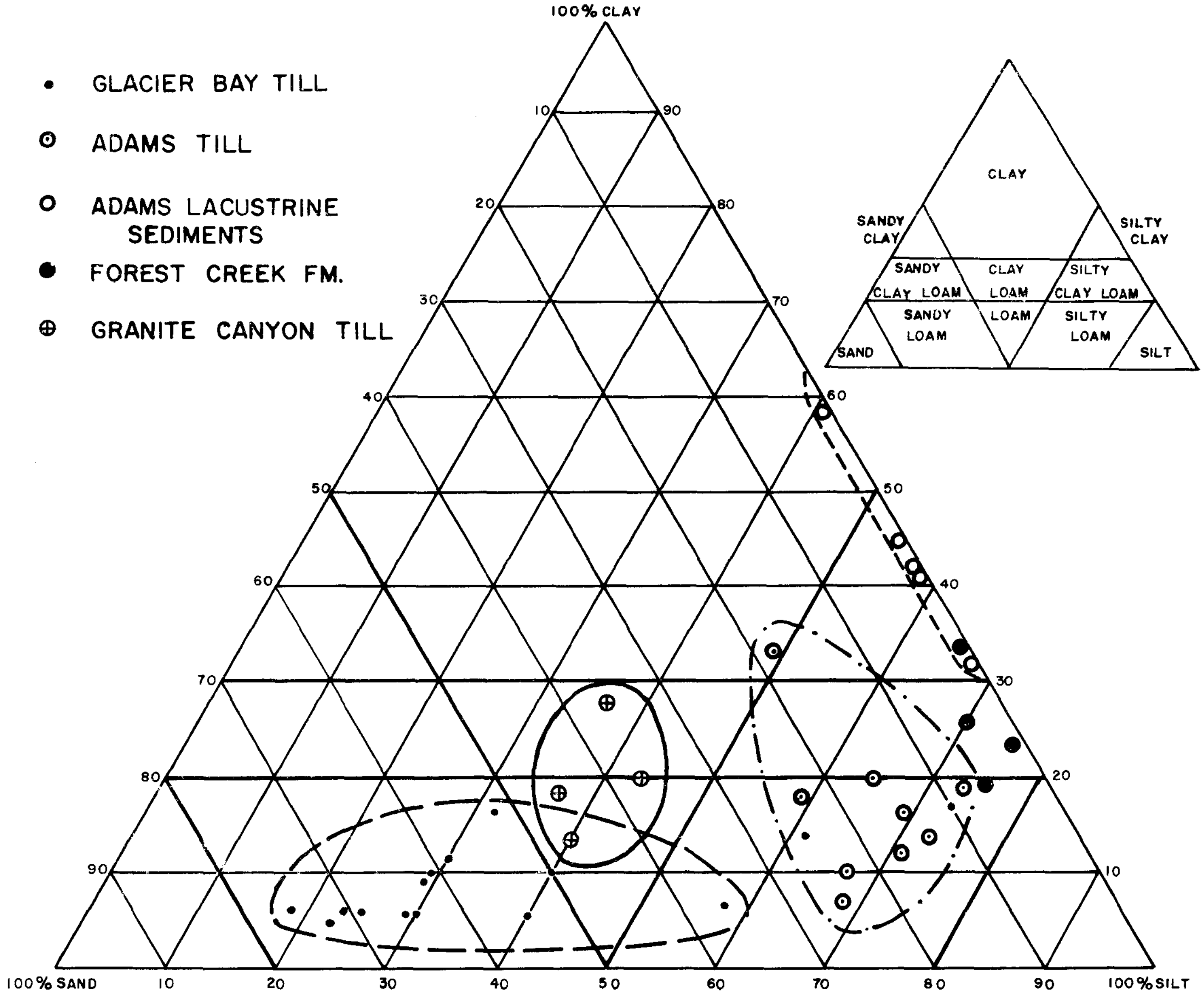


Figure 8. Mechanical composition of tills and fine-grained sediments.

Table 1. Mechanical composition of formations

Formation	Statistic[a]	Sand	Silt	Clay
Granite Canyon Till	X	41.0	39.2	19.8
	V	27.3	8.3	29.6
	N	4	4	4
Forest Creek Silt and Clay	X	2.8	71.7	25.5
	V	7.6	21.6	40.3
	N	4	4	4
Adams Silt and Clay	X	0.0	56.2	43.8
	V	0.0	95.7	95.7
	N	5	5	5
Adams Till	X	18.5	66.2	15.3
	V	24.5	53.5	60.4
	N	10	10	10
Glacier Bay Till[b] (Muir Inlet)	X	54.2	36.1	9.7
	V	30.6	15.8	10.7
	N	25	25	25
Glacier Bay Till (Adams Inlet)	X	55.3	35.9	8.8
	V	347.0	250.0	17.0
	N	15	15	15

[a]X, mean value in percent; V, variance; N, number of samples
[b]Data from Haselton, 1967

Table 2. Pebble lithologies of tills and gravels

Formation & Location	Number of Samples	Percentage of Rock Types: Dike Rock	Volcanics	Plutonic Igneous	Limestone	Graywacke	Hornfels	$\frac{PI}{M}$[a]	Var.[b]
Granite Canyon Till	6							0.02	3.8×10^{-4}
Goddess valley	2	13	7	1	15	59	5	0.01	
Granite Canyon	4	25	11	1	2	45	16	0.02	
Van Horn Gravel	20							0.39	0.35
Adams Inlet	7	37	11	5	8	33	6	0.11	
Adams Island	5	36	13	22	5	19	5	0.76	
Adams Valley	3	22	12	--	21	38	7	0.01	
Goddess valley	2	24	20	11	12	31	2	0.24	
Granite Canyon	1	53	6	26	--	9	6	1.7	
White Valley	2	2	54	--	16	21	7	0.00	
Adams Till	19							0.80	0.14
Adams Inlet	13	24	10	27	5	17	17	0.69	
Adams Valley	4	22	11	37	3	14	13	1.2	
Goddess valley[c]	2	30	6	16	1	38	9	0.33	
Berg Gravel	25							0.83	0.31
Adams Inlet	14	25	7	29	8	21	10	0.74	
Adams Valley	2	30	3	30	9	13	15	0.81	
Goddess valley	5	30	12	15	8	24	11	0.35	
Granite Canyon	4	32	10	27	2	11	18	0.87	
Glacier Bay Till	37							0.70	0.32
Adams Inlet	20	30	8	19	10	17	16	0.44	
Adams Valley	6	31	8	24	6	14	17	0.65	
Goddess valley	5	29	12	15	3	31	10	0.34	
Granite Canyon	6	35	10	29	2	11	13	1.1	
Seal River Gravel	16							2.3	11.0
Adams Inlet	1	22	9	19	5	31	14	0.38	
Goddess valley	7	19	14	16	7	35	9	0.31	
Granite Canyon	2	28	8	52	--	6	6	4.3	
Seal Valley	6	40	10	36	2	8	4	2.5	

[a]Ratio of plutonic igneous to metasedimentary & sedimentary rocks, average of all samples in the formation
[b]Variance of PI/M ratios
[c]One sample from rhythmic clay

The clay mineral analyses of eight samples (Table 3) from several formations in the map area show similar composition with illite dominant or a major component along with chlorite. Total expandable clays are less than 1.5 parts per ten, and quartz is less than 2.5 parts per ten in all samples.

The limited number of samples and the semiquantitative method of determining clay mineral percentages preclude using clay minerals as a very reliable basis for differentiation and correlation of formations in Adams Inlet. There are minor differences, and some close similarities, in the samples and these might be significant (Table 2). In the discussions of the clay minerals these differences and similarities have been emphasized, but the limits of the method should be borne in mind in evaluating the conclusions. Clay mineral analyses are probably most useful in this study in illustrating changes in the till on weathering. This is the case with the Granite Canyon till, from which a weathered (66-8) and an unweathered (66-9) sample were analyzed.

In interpreting the X-ray diffractograms of the clays (Appendix) the 10 Å peak is considered to be the basal (001) reflection of illite. The sharpness of this peak on all samples suggests that micaceous material is well crystallized. Part of the air-dry 14 Å peak is montmorillonite because the samples treated with ethylene glycol show a 17 Å peak. The 14 Å peak decreases on heating to 400°C due to a collapse of montmorillonite to 10 Å. Slightly lower 14 Å peaks of some heat-treated (400°C) samples, compared with the ethylene glycolated samples, suggest the presence of minor amounts of vermiculite. Heating to 550°C results in the destruction of the 3.5 and 7 Å peaks and enhancement of the 14 Å component. This suggests an iron-rich chlorite; however, the 3.5 and 7 Å (001) peaks of kaolinite would also be destroyed by such treatment. Several samples (58-8, 67D-8, 66-9) were leached with HCl to destroy the chlorite and were rerun to check for kaolinite. No kaolinite peaks were present. In the diffractograms of these leached samples (not shown) there were increases in the heights of illite, quartz, and feldspar peaks in the air-dry samples concomitant with the loss of chlorite. Ethylene glycolated and heat-treated tests were not made on these samples and therefore it is not known what effect leaching may have had on the interstratified and expandable clays.

There is also a broad 14 - 17 Å plateau or peak in the ethylene glycolated diffractograms of all samples. This suggests an interstratified clay, probably chlorite and montmorillonite. Quartz abundance was also estimated in the clay-size fraction, as were the amounts of amphiboles and interstratified 10-14 Å clay. The interstratified 10 - 14 Å is probably a random mixture of illite and montmorillonite, possibly with some vermiculite.

The relative youthfulness of the deposits in Adams Inlet is evident in the clay-sized mineral suites. Chlorite is abundant and illite is well crystallized. Amphiboles occur in some of the deposits; such

Table 3. Relative abundance of clay minerals in formations

Formation and Sample*	Illite	Chlorite	Interstratified 14 - 17 Å	Montmorillonite	Total Expandable Clays	Quartz	Interstratified 10 - 14 Å	Amphibole
Glacier Bay Till								
66 - 2	5	3	0.5	-	0.5	1.5	-	+
74 - 1	6	3	Tr	-	Tr	1	-	-
Adams Till								
58 - 6	4	4	0.5	0.5	1	1	+	+
Adams Silt and Clay								
58 - 8	4	4	1	Tr	1	1	+	+
Van Horn Diamicton								
66 - 6	4.5	4.5	0.5	-	0.5	0.5	+	+
Forest Creek Silt and Clay								
67D - 8	4.5	3.5	1.5	-	1.5	0.5	+	+
Granite Canyon Till, Weathered								
66 - 8	8	1	0.5	Tr	0.5	0.5	-	-
Granite Canyon Till, Unweathered								
66 - 9	5	2.5	Tr	-	Tr	2.5	-	-

*Data are given in parts per ten; minus sign indicates not detected; plus sign indicates present but abundance not determined; Tr = Trace, indicating 2 percent or less.

minerals are uncommon in weathered materials (Jackson and others, 1948).

The clay mineral composition of the two samples from the Granite Canyon till is given in Table 3. Sample 66-8 from the weathered upper 0.5 m of the 3-m-thick till is mostly illite. Weathering, as is evident from a comparison with the composition of the parent material (66-9), has caused a decrease in chlorite and quartz, and an increase in interstratified 14 - 17 Å material, montmorillonite and illite. Some of the expandable material may be forming from illite and ferromagnesium minerals (Thorp and others, 1959), and possibly from chlorite. Some expandable material was present in the original material and its relative abundance may have increased with weathering. Since a detailed study employing bulk densities, thin sections, particle-size analyses, and mineralogic analyses of various size fractions was not carried out, little information on the soil-forming processes was obtained. Sample 66-8, from the B horizon of the soil, has lost the more mobile constituents of the original material and gained others by illuviation. Montmorillonite, which is usually among the finest of the clay minerals, has probably migrated to the lower part of the B horizon. The reduction of quartz to one-fifth of the original amount in this size fraction suggests intense weathering, although much silica can be lost during the early stages of development during depletion of carbonates (Wright and others, 1959). The increase in illite may be due to the reduction in amount of chlorite and quartz and not an absolute change in the amount of illite.

Partial elemental analyses (Table 4) were done on the sand and silt fractions of samples from formations in Adams Inlet in an effort to determine differences between tills, differences in deposits of the same till from different valleys, and degree of weathering in one till and a diamicton. Samples were also leached with HCl in an effort to further determine differences and similarities of deposits. Selection of elements to be analyzed depended in part on available standards. In all leached samples Ca, Fe, and Mn decreased. The elements Al, K, Ti, and Zr showed relative increases in almost all samples tested. Brief discussions of the results of the elemental analyses are given in the stratigraphic description of each formation. In comparing the elemental composition of the tills, differences of 10 to 15 percent in the values for the elements (Ti, Zr, K) of stable minerals are considered significant enough to differentiate the tills (Beavers, 1960; Smeck and others, 1968).

Comparison of the two samples (66-8, 66-9) of Granite Canyon till (Table 4) reveals a reduction in Ca content as a result of weathering. This is probably due to weathering out of carbonates and possibly some of the Ca-feldspar. The average calcite content of two unweathered samples of this till is 7.0 percent (Table 5); calcite is probably absent in the weathered sample (66-8). Zr is the only other element that decreases in the weathered sample and this is unusual because of the stability of zircon. The expected increase is seen in the HCl-treated samples. The ratio of zirconium to calcium has been used to indicate

Table 4. Partial elemental analyses of sand and silt fraction of samples analyzed for clay minerals

Formation and Sample	Percent						Counts per sec.
	Ti	Zr	Fe	Mn	K	Ca	Al
Glacier Bay Till							
66 - 2	0.227	0.174	3.06	0.079	1.34	4.9[b]	586[c]
66 - 2 (HCl)[a]	0.268	0.192	2.36	0.064	1.47	3.2	687
74 - 1	0.240	0.176	3.35	0.087	1.37	5.0	595
74 - 1 (HCl)	0.287	0.206	2.49	0.065	1.49	2.5	654
Adams Till							
58 - 6	0.262	0.172	3.59	0.101	1.15	5.1	614
58 - 6 (HCl)	0.305	0.169	2.58	0.085	1.30	3.9	674
Adams Silt and Clay							
58 - 8	0.308	0.226	4.16	0.127	1.58	5.0	807
58 - 8 (HCl)	0.368	0.216	2.49	0.080	1.73	3.4	839
Van Horn Diamicton							
66 - 6	0.308	0.212	4.01	0.156	1.66	2.2	725
66 - 6 (HCl)	0.299	0.207	2.21	0.071	1.67	2.0	723
Forest Creek Clay and Silt							
67D - 8	0.340	0.184	4.61	0.146	1.67	5.5	725
67D - 8 (HCl)	0.437	0.171	3.29	0.100	1.80	2.2	783
Forest Creek Ash							
67D - 6	0.176	0.142	2.85	0.077	1.64	2.0	784
Granite Canyon Till, Weathered							
66 - 8	0.317	0.166	4.05	0.095	1.76	1.08	667
66 - 8 (HCl)	0.322	0.188	2.07	0.044	1.87	0.84	727
Granite Canyon Till, Unweathered							
66 - 9	0.244	0.186	3.53	0.080	1.55	4.8	571
66 - 9 (HCl)	0.315	0.179	2.14	0.047	1.76	1.04	708

[a] Leached with HCl to remove carbonates
[b] Because of too few standards and non-linearity of the resulting curve the error for values > 1.7 is ± 0.2; otherwise the error is ± 0.05
[c] No standards available

degree of weathering (Wright and others, 1959). Calcium carbonate is leached out early in the process of weathering and a lower ratio indicates less weathering. For sample 66-8 (Table 4) the Zr/Ca ratio is 0.154, and for 66-9, the unweathered till, the ratio is 0.038.

Table 5. Carbonate Analyses of the -200 Mesh Fraction

Formation	No. of Samples	Dolomite (Wt. %)	Calcite (Wt. %)	Total Carbonate	Ratio Cal./Dol.
Glacier Bay Till	3	2.2	5.6	7.8	2.5
Adams Till	3	1.3	8.6	9.9	6.6
Adams Silt and Clay	3	1.0	9.4	10.4	9.4
Van Horn Diamicton	1	0.3	0.1	0.4	0.3
Forest Creek Silt and Clay	3	1.7	7.8	9.5	4.6
Granite Canyon Till	2	6.6	7.0	13.6	1.1

The other elements (Table 4) of the weathered layer of Granite Canyon till show an increase in abundance over the unweathered unit, probably due both to the decrease in carbonate and to the illuviation of the elements. In comparing the HCl-leached samples (66-8, 66-9), there is a slight decrease in Ca, probably plagioclase but possibly some carbonate, and decreases of Fe and Mn in the weathered till. The expected increases in Ti, Zr, and K, elements of more resistant minerals, do occur in the weathered till.

This 50-cm-thick weathered till unit apparently developed between the beginning of the Holocene and the end of the Hypsithermal. It is overlain by 60 cm of gravel with talus blocks, which in turn is overlain by a forest bed, dated at 3,850 years B.P. (I-3068). Ground water may have caused further weathering of the till unit after burial of the forest bed. Movement of ground water would be facilitated by the 60-cm layer of gravel over the till, and organic acids that promote weathering would also be readily available.

Glacier Movements During the Wisconsin Glaciation

Till Fabric

Till fabric measurements on Granite Canyon till were made at two localities, one on Granite River and the other on Goddess river. The results are given on stereonets (Fig. 9) and on the directional features map (Fig. 10). At Granite Canyon (Station 67-60) the fabric, from about

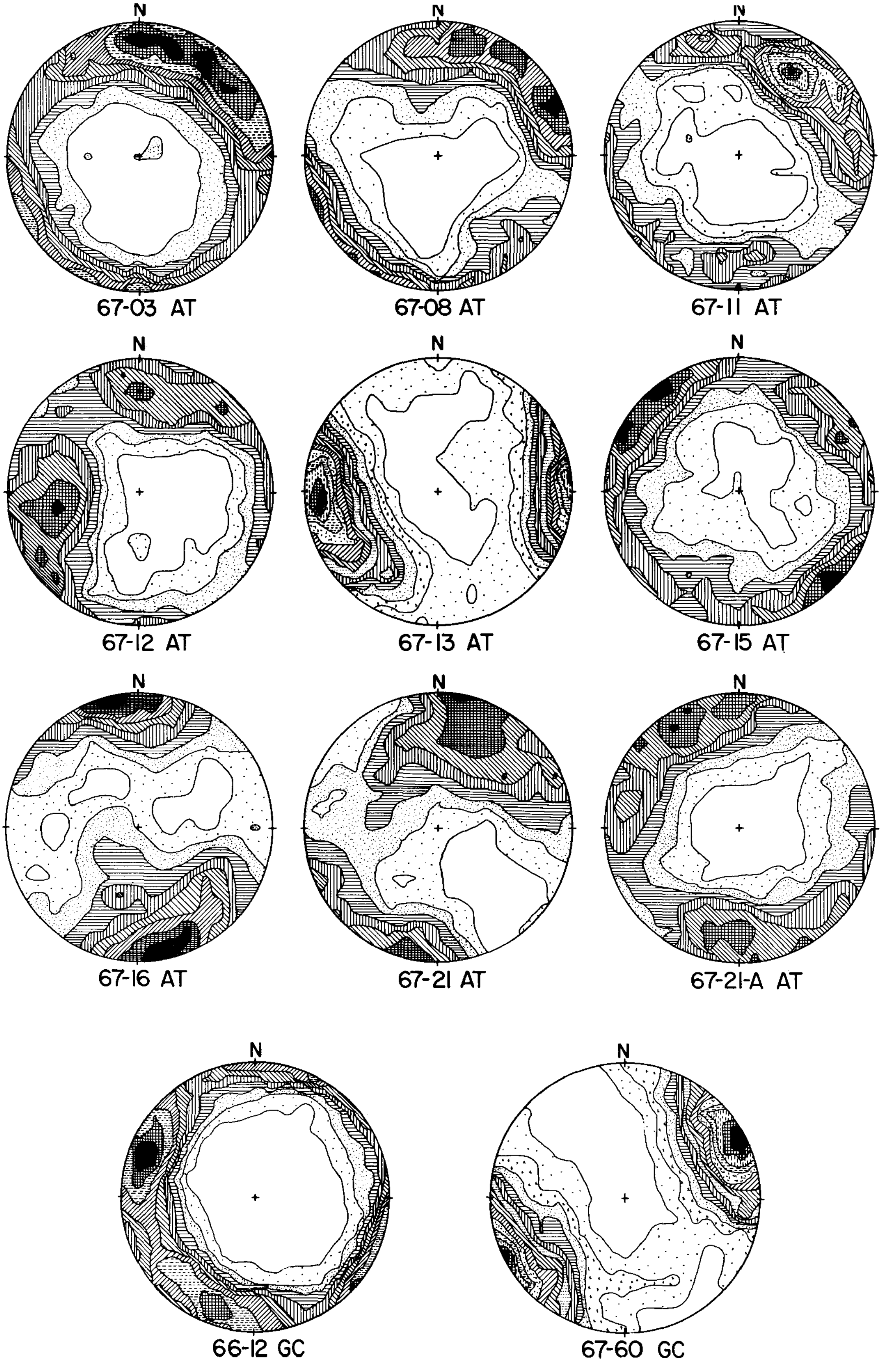


Figure 9. Till fabric diagrams of Adams till (AT) and Granite Canyon till (GC). White areas are the lowest concentrations. Contour interval is one sigma, contoured according to the method of Kamb (1959). See text for locations.

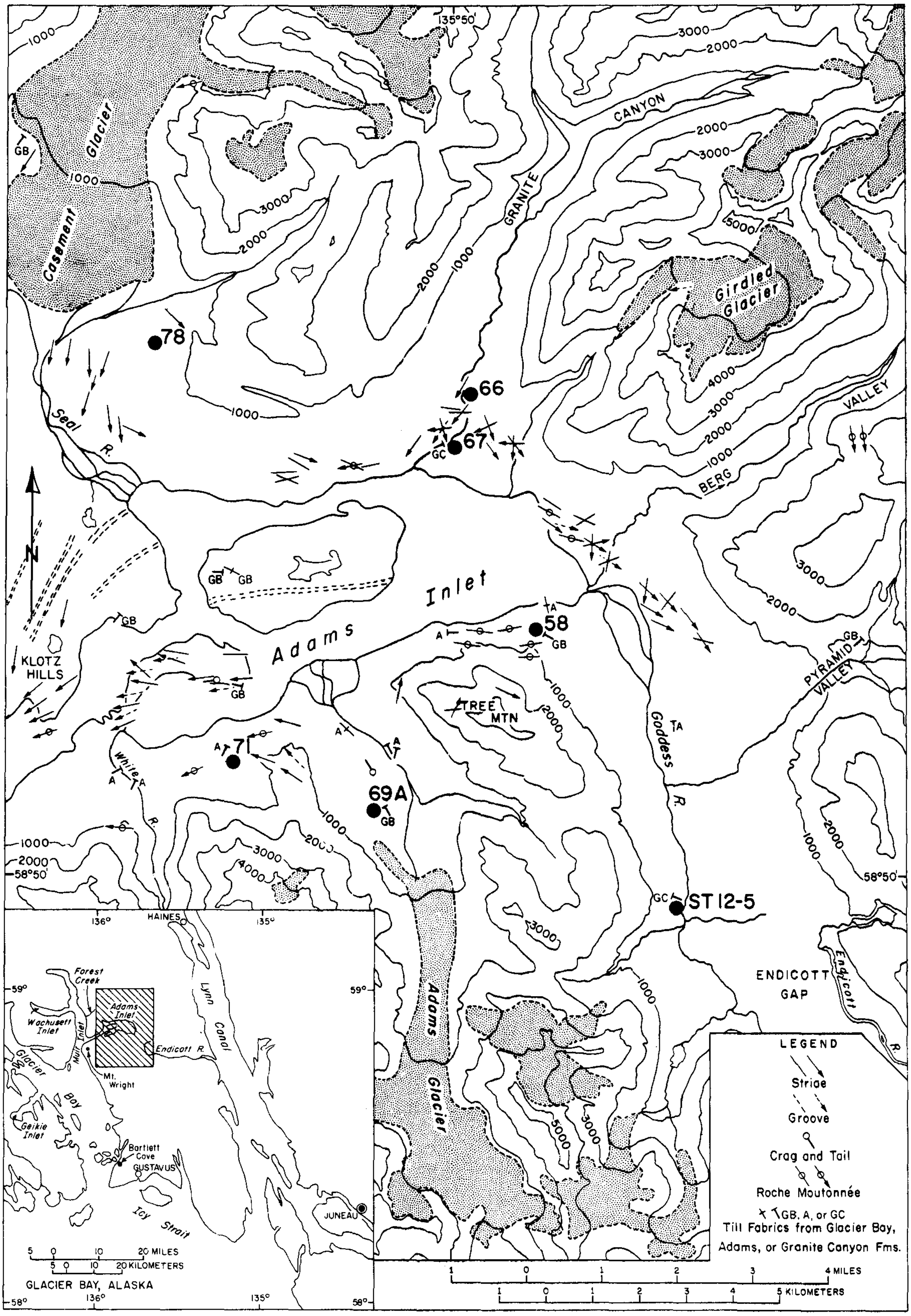


Figure 10. Directional features map of Adams Inlet area. Parallel broken lines are medial moraines. Fabric symbols: bar on end indicates direction from which ice came; bar in middle indicates direction unknown.

1.5 m to 2.0 m above the bedrock, showed the highest concentration of points at N 62° E with a plunge of approximately 5°. This is the direction from which the glacier came, because the long axes of pebbles are imbricate in the up-glacier direction (Harrison, 1957). This result agrees well with the N 65° E direction of striae and grooves on the bedrock, and indicates a flow out of Granite Canyon.

At Goddess river the main maximum was at N 70° W with a transverse maximum at S 20° W. The glacier flow from N 70° W would come down from the nearby mountain south of Tree Mountain, and head toward Lynn Canal. Proximity to the mountain and bedrock knobs that probably affected the ice-flow direction may have resulted in a local variation of the regional pattern.

Streamline Hills and Roches Moutonnées

Ice-sculptured bedrock in the form of streamline hills and large grooves exist on the northeast side of Tree Mountain (Fig. 11). Long axes of the asymmetrical hills and bosses trend southeastward around the mountain. The flow direction is uncertain, but was probably southward through Endicott Gap. It is inferred that the bedrock features were formed during Wisconsin Glaciation because the relief of up to 50 m suggests a longer period of glacial erosion than that available to Neoglacial ice. An 850-m-long streamline limestone hill at the east end of Adams Inlet was probably formed in large part by Wisconsin ice; however, it was no doubt modified by Neoglacial ice. The small (2 m high) roche moutonnée on the southeast end was probably the result of Neoglacial ice movements.

A large roche moutonnée occurs on the south side of Adams Inlet opposite Camp Adams. Flow was to the west. Two roches moutonnées occur at an elevation of 850 m (2800 ft) on the east side of the map area, 3.5 km southeast of Mt. Kimber (Fig. 10). Here, and probably over Tree Mountain, Wisconsin ice flowed southward toward Endicott Gap. Another component of ice flow was westward out of the entrance to Adams Inlet. Another large roche moutonnée, known locally as "Ripple rock" (Fig. 12), occurs near the north side of the entrance to Adams Inlet. It is exposed only at low tide and is a menace to navigation. Wisconsin ice may have had a large part in formation of this feature.

Age and Origin

Wood from a peat layer on top of the overlying Forest Creek Formation exposed at Section 67 in Granite Canyon was dated by radiocarbon at 10,940 ± 155 years B.P. (I-2395) (Table 6). A spruce cone collected by R. P. Goldthwait and the writer from 60 cm below the top of the type section of the Forest Creek Formation yielded an age of 11,170 ± 225 years B.P. (I-2396), which overlaps the date from on top of the

Figure 11. Streamline hills on the east side of Tree Mountain. Note the southward-sloping trim line on the north side of cirque. View southwest from Berg Creek.

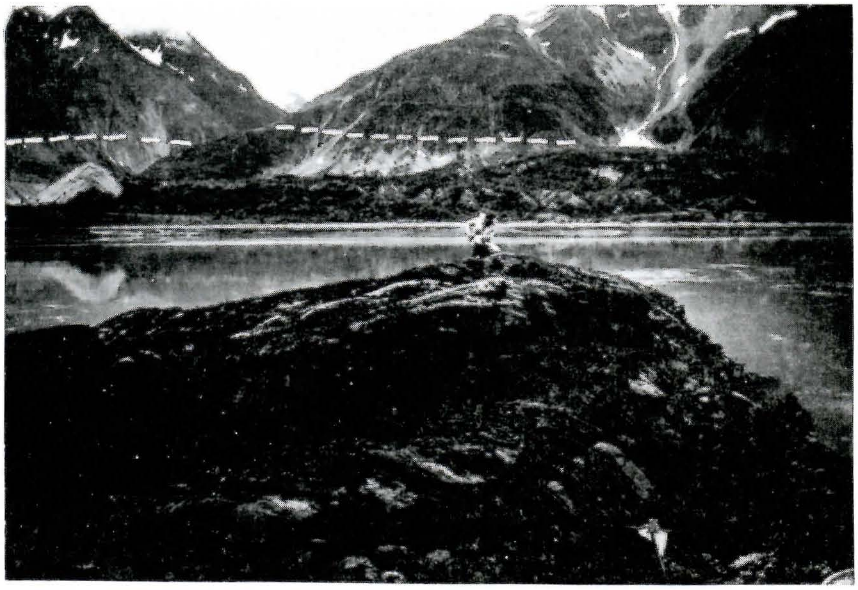


Figure 12. "Ripple rock", a roche moutonnée, at the entrance to Adams Inlet. View is southward toward White Valley at a low tide of -1 m (-3.4 feet). Dotted line is approximate position of ice in a photograph taken by H. F. Reid in 1890. Kame at entrance to valley is 90 m high. Striae bearing N 68° W on top of "Ripple rock" indicate glacier flow was from left to right out of Adams Inlet.

Table 6. Radiocarbon dates from Adams Inlet and surrounding area

Sample number	C^{14} Date BP BC/AD	Location and Material	Elevation (meters)	Collector
		Forest Creek Formation		
I-2396	11,170 ± 225 BP 9,220 BC	Forest Creek Gorge Section. Spruce Cone.	30	McKenzie Goldthwait
I-2395	10,940 ± 155 BP 8,990 BC	Section 67, Granite Canyon. Peat on top of formation.	32	McKenzie
I-1615	10,400 ± 260 BP 8,540 BC	Upper Forest Creek. Wood.	60	Haselton
I-1303	10,000 ± 220 BP 8,050 BC	Forest Creek Gorge. Shells.	27	Haselton
		Van Horn Formation		
I-3068	3,850 ± 110 BP 1,900 BC	Section 66, Granite Canyon. Twig in peat at base of Van Horn diamicton.	89	McKenzie
I-164	3,710 ± 110 BP 1,760 BC	South Shore Adams Inlet, east end, 60 cm diameter log.	3	Miller[a]
I-3398	2,390 ± 110 BP 440 BC	West side Muir Inlet near U.S.C.G.S. Station "Quill." Allochthonous wood at base of clay.	8	Goldthwait[b]
I-2302	2,190 ± 105 BP 240 BC	Near Casement Base. Stump on bedrock.	300	Goldthwait
I-2394	1,980 ± 100 BP 30 BC	South shore Adams Inlet, east end. Autochthonous stump.	19	McKenzie
I-85	1,975 ± 150 BP 25 BC	South side, east end of Adams Island.	10	Goldthwait
		Adams Formation		
I-3069S	1,770 ± 100 BP 180 AD	Section 66, Granite Canyon. Allochthonous (?) wood at base.	90	McKenzie
I-2687	1,700 ± 100 BP 250 AD	Section 58, south side, east end Adams Inlet. Autochthonous stump.	24	McKenzie
I-2300	1,690 ± 100 BP 260 AD	Near Section 71, south side Adams Inlet. Allochthonous wood near bedrock.	125	Goldthwait
I-2301	1,560 ± 95 BP 390 AD	Near Section 71, south side Adams Inlet. Allochthonous wood in till.	55	Goldthwait
I-3151	1,535 ± 100 BP 415 AD	Section 78, gorge on southeast side of Casement Glacier. Branch in peat on till.	120	McKenzie
		Berg Formation		
I-3150	1,750 ± 100 BP 200 AD	Section 69A, Adams mesa. Allochthonous wood 5 m below Glacier Bay Till.	230	McKenzie

[a]D. J. Miller of the U.S.G.S. Sample submitted by R. P. Goldthwait.
[b]Submitted by McKenzie.

Forest Creek Formation. These dates give a close minimum age for the Granite Canyon till because the thickness of the Forest Creek Formation is not great (about 2 m) and because of the absence of a weathered contact with the till. As the ice that deposited the till retreated, the sea reached elevations of 28 m in Granite Canyon and 59 m in Forest Creek (Haselton, 1966). The deposit on Goddess river lies beneath undated organic remains at the base of the Adams Formation; there the base of the Granite Canyon till is not exposed. The Wisconsin age Granite Canyon Formation is the oldest known Pleistocene deposit in Glacier Bay.

Correlation

The Granite Canyon till is probably equivalent in age to some deposits of Muir Formation. Haselton (1966) describes the Muir till as bluish-gray sandy loam with few boulders and cobbles, and yellowish-brown or reddish-brown in the oxidized parts. It crops out in 5 places on the east side of Muir Inlet, and in 3 places on the north side of Wachusett Inlet (Goldthwait and others, 1966). Two occurrences on the south side of Adams Inlet, believed by Goldthwait to be Muir till, have been shown by stratigraphic correlation and radiocarbon dates to be part of the Adams Formation.

A diamicton[1] in the same stratigraphic position as the Granite Canyon till was discovered by the writer at the type section of the Forest Creek Formation. Although diamictons may form from debris dropped by icebergs, ice shelves, and floating glaciers into massive deposits of silt and clay, these deposits are glaciomarine (water-laid till, marine till, or pebbly marine clay) and not true tills. Use of the term till implies deposition directly from a glacier or incorporation and redeposition of debris by a glacier. It is not always possible to determine to what extent, if any, a glacier deposited material in a marine or lacustrine environment; undisturbed remains of aquatic life in such material, however, preclude use of the term till. Similar deposits without shells may be tills although it is conceivable that they were formed far beneath a floating glacier where life could not exist. The purplish-gray pebbly-clay diamicton at Forest Creek is probably a till because of the absence of readily visible indicators of marine life (shells, barnacle plates) and the high sand content--45 percent, which is comparable to other samples of Wisconsin till at Granite Canyon and which is coarser than the Forest Creek Formation. Only 30 cm of the diamicton were exposed, and the contact with the pebbly, shell-bearing clay of the Forest Creek Formation was gradational. Although bedrock was not seen below this material, the bedrock beneath the Forest Creek Formation is striated (Haselton, 1966).

[1]Flint and others (1960): a diamicton is a terrigenous unconsolidated deposit that contains a wide range of particle sizes.

Haselton (1966) and Goldthwait (Goldthwait and others, 1966) have assigned a Late Wisconsin age to the Muir Formation. At one exposure in Muir Inlet and two in Wachusett Inlet the Muir Formation rests on bedrock; at other places the base of the till is not exposed except in Upper Forest Creek where Muir till overlies the poorly exposed Forest Creek Formation. This section is the key to the position of the Muir till because in other sections it is the lowest exposed deposit. A radiocarbon-dated log at the interface of the Forest Creek Formation and the Muir till has an age of 10,400 ± 260 years B.P. (I-1615) (Table 6). The stratigraphic position and this basal date certainly place this occurrence of the Muir till as post-Forest Creek; however, the possibility exists that other deposits of the till are pre-Forest Creek and thus equivalent to the Granite Canyon till.

Lithologic, mineralogic, mechanical, and fabric analyses can be used for correlation purposes if the source areas and the degree of weathering, if any, are the same. In the Muir Inlet area, Haselton's Muir till shows marked differences in lithologic composition between Forest Creek and other localities. No mineralogic or mechanical analyses were made on the Forest Creek exposure of Muir till, and fabric analyses would probably not be useful for correlation purposes in this mountainous area. In attempting to correlate the other Wisconsin tills in Muir Inlet and those in Adams Inlet, pebble lithologies fail. No mineralogic analyses are available from either area. The mechanical composition of the Wisconsin till beneath the Forest Creek Formation at Forest Creek is similar to those of the Granite Canyon till in Adams Inlet (Fig. 8). The other Wisconsin tills (Muir) in Muir Inlet are coarser than those in Adams Inlet.

Stratigraphic position and absolute dates are the only ways of relating these deposits of Wisconsin till. The only evidence that can be offered for the equivalency of most Muir tills (except at Forest Creek) and Granite Canyon tills is their lowermost position in stratigraphic sections.

If we assume that most deposits of Muir till are equivalent to the Granite Canyon till (Wisconsin) and that the Muir till at Forest Creek is an isolated occurrence, then there are three possible interpretations for this Forest Creek Muir till: (1) it represents a major advance of ice in Glacier Bay, (2) it represents a minor advance of ice in the vicinity of Casement Glacier, and (3) it is a flow till or solifluction deposit. The first possibility is unlikely since no other tills in Glacier Bay have been found in the same stratigraphic position. Elsewhere the Forest Creek Formation is overlain by lacustrine clays, gravels of uncertain age, or solifluction-talus material. At the Casement locality (Goldthwait and others, 1966) the stratigraphic position of the Forest Creek Formation is uncertain. A glacial advance in the vicinity of Forest Creek at this time would be bracketed by a date of 10,400 years B.P. (I-1615) (Table 6) on a log at the top of the Forest Creek Formation, and a date of 7,025 years B.P. (I-84) on allochthonous wood in gravel stratigraphically above the Muir till. There is evidence in other areas for a glacial advance during this interval, or for the persistence of Wisconsin ice until 8,000 or 9,000 years ago. In northern Alaska the Anaktuvuk Pass advance is suggested to have been between 8,000 and 9,000 years ago (Porter, 1966), and at Snoqualmie Pass, Washington, Wisconsin ice persisted until shortly before 7,200 years B.P. (UW-73) (Porter and Denton, 1967). Other dates and evidence for persistence of ice or readvances at this time are given by Goldthwait (1966). Depending on the locality there is some evidence to support a major advance or persistence of Wisconsin ice with possible minor fluctuations.

The second possibility, a minor advance, could be either a minor pulsation of retreating Wisconsin ice or a surge from a nearby previously stable valley glacier. Pulsation of an ice margin near this locality would result in bracketing of the Forest Creek Formation by Wisconsin (Muir) till. If the pulsation were small enough no similar sequence would be expected to the west because that area would still be covered by ice; and to the southeast in Adams Inlet the sequence would be absent because the ice did not advance to that point. An advance of Casement Glacier would be unlikely without general advance of most glaciers in the Monument because the Casement Glacier is in the eastern part of the Monument where precipitation is lower and the catchment basins smaller. The deposit of Muir till could also be the result of a surging Casement Glacier, and the east-west till fabric (Haselton, 1966) may suggest this source. The absence of a soil profile below the Forest Creek exposure of Muir till, although possibly due to erosion, suggests a short time interval before burial of the underlying Forest Creek Formation at Forest Creek. The dated log (10,400 years B.P.) at the interface of the Forest Creek clay and the Muir till gives a maximum age for the Muir till close to the age of the Forest Creek Formation.

The third possibility, that of some type of mass movement, is difficult to prove or disprove. The nature of the till fabric may be one method of determining whether or not the deposit was a flow till or the result of solifluction (Harrison, 1957); however, there is no indication of an unusual fabric at this location (Haselton, 1966). Lithologic differences may be helpful in some cases in determining influx of material of a solifluction deposit, but a flow till from a glacier carrying a lithologically uniform load would probably be little, if any, different from a normal till. Solifluction of a deposit of Wisconsin till, weathered during the deposition of the Forest Creek Formation and for several hundred years thereafter, could result in the Muir till over the Forest Creek Formation at Forest Creek and could also explain the low carbonate content of this till.

From the above discussion it can be seen that the Granite Canyon Formation may be correlative with part of the Muir Formation as defined by Haselton (1966). There is no definite evidence that the till overlying the Forest Creek Formation at Upper Forest Creek is equivalent to the other deposits of Wisconsin till, or that this possible post-Forest Creek advance was widespread. The Forest Creek deposit of Muir till may represent a minor pulsation or readvance during the Wisconsin retreat, or it may be a solifluction deposit or flow till.

Forest Creek Formation

Definition and Type Section

The Forest Creek Formation was defined by Haselton (1966) as a bluish-gray to gray thin-bedded fossiliferous marine clay and silt. The Forest Creek Formation, in the type section at a bedrock gorge on Forest Creek, 2 km from the mouth of the creek, rests on striated bedrock and is overlain by a fine- to medium-grained thin-bedded sand that Haselton suggests is a beach deposit. This deposit is covered by solifluction material and recent talus.

In the type section, abundant striated pebbles, suggesting a nearby ice margin, are covered on the upper surface with barnacle plates. Fossils, identified by A. La Rocque, include gastropods, *Colus spitzbergensis*, *Trichotropis borealis*, and *Neptunea lyrata*; and the pelecypods *Trachycardium quadragenarium*, *Mya arenaria*, *Macoma* sp., *Hiatella arctica*, and *Chlamys islandicus*.

Distribution

The Forest Creek Formation is exposed in four localities other than the type section. These are in upper Forest Creek (59 m elevation), south side of Casement Glacier (30 m), Granite Canyon (28 m), and at the terminus of Reid Glacier (3 m). These few deposits are 1 m to 2 m thick and, except for the deposit in Granite Canyon which overlies the Granite Canyon till, they rest on bedrock. The exposures in Granite Canyon occur on both sides of the river over a distance of 200 m; their maximum thickness is 2.3 m. On the south side of Adams Inlet at about 25 m elevation near Section 70 (Fig. 2), shells and shell fragments occur in slump of green silt and clay. This is probably the Forest Creek Formation but the section was too highly slumped to determine the stratigraphic relationships.

Nature of Deposits

The deposit in Granite Canyon is a gray laminated clay and silt, with black mottles, that appears massive in section. It readily oxidizes to grayish green. A few striated pebbles, some with barnacle plates, and numerous pelecypods and gastropods are present. A 0.5-m-thick layer of volcanic ash occurs about 1.3 m below the top of the Forest Creek Formation at Granite Canyon.

Grain Size Distribution

Four mechanical analyses (Fig. 8, Table 1) were run on samples from the Forest Creek Formation in Granite Canyon. The average of these analyses is 2.8 percent sand, 71.7 percent silt, and 25.5 percent clay. No mechanical analyses are available for the Forest Creek Formation at the type section; however, inspection showed the exposure to contain numerous pebbles, which suggest the possibility of a higher percentage of sand as well. Pebble counts are not available but it is inferred that the percentages of pebble types in the Forest Creek Formation would be similar to those in the underlying Granite Canyon till.

Clay Minerals

Illite is the dominant clay mineral in the Forest Creek Formation; chlorite is next in abundance (Table 3). The suite of clay minerals in the Forest Creek Formation, and also in the tills of the map area, is similar to that in modern marine sediments in this part of the world (Griffen and Goldberg, 1963). Compared with the composition of the underlying Granite Canyon till, there is less illite and quartz, but more chlorite and expandable clays in the Forest Creek Formation. The differences, however, are not great.

The question arises whether or not these differences between clays of the Forest Creek Formation and the Granite Canyon till are a result of alteration, synthesis, diagenesis of clay minerals in the ocean, or physical factors. Distribution patterns of clay minerals in modern environments have been used to support alteration of detrital clay minerals in saline environments (Grim and Johns, 1954; Powers, 1957; and others). Diagenetic changes have been supported on a theoretical basis by Keller (1963) using data of Carroll and Starkey (1959), and others on the reactions of clay minerals with cations in sea water. Synthesis of clay minerals from amorphous aluminosilicates entering the oceans has been postulated by Mackenzie and Garrels (1966). Although it has been demonstrated that alteration of clay minerals in a marine environment does occur, these changes are thought by many (Weaver, 1959, Griffen, 1962; and Biscaye, 1965) to be of secondary importance. Weaver (1959) has suggested that distributions of modern sediments may be explained by preferential flocculation, current sorting, different source areas, floods, and periodic variation in composition and concentration of river detritus.

The mineral composition of the Forest Creek Formation might be expected to be similar to that of the older Granite Canyon till in the same area, if the marine sediments were derived from the glacier that deposited the till (rafting on icebergs), and from recently exposed and easily eroded surfaces of till that no doubt blanketed the valley floors and walls. The differences in the samples of glaciomarine material and clay of the Granite Canyon till may be due to one or more of the physical factors suggested by Weaver (1959); they may also indicate different source areas. Increase in chlorite (Table 3) due to alteration of the minerals has been suggested (Grim and Johns, 1954; and others) but was not supported by Biscaye (1965). Insufficient data are available in the case of the Forest Creek Formation to support or reject this origin for the chlorite. Increase in montmorillonite due to weathering of volcanic ash that might occur in the silt and clay is not probable because even in the ash layer the glass shows no weathering and montmorillonite is absent.

Elemental and Carbonate Analyses

Compared with the composition of the unweathered Granite Canyon till, elemental analyses of the Forest Creek Formation (Table 4) show a higher percentage of all elements tested except zirconium, which was about the same in both cases. Amounts of the same elements differ by as much as one-third and suggest that the source of the sediment being deposited in one area of Granite Canyon during Forest Creek time was not restricted to the Granite Canyon Formation either as derived from erosion of the till blanket from surrounding slopes or ice-rafting of debris from nearby receding glaciers. The slightly higher Ca content in the Forest Creek Formation may correspond to the greater percentage of calcite (Table 5) in that unit. The dolomite percentage for this formation is much lower than that of the Granite Canyon till. The overall results of the carbonate analyses show the Forest Creek Formation to be most similar to the Adams till.

Fossil Assemblage

Many of the fossils from the Forest Creek Formation of Granite Canyon are of the same genera as those in the type section (Haselton, 1966; and listed earlier). They were identified by A. La Rocque, who provided the environmental interpretation, and consist of the following:

Bivalvia (Pelecypoda):

Chlamys islandicus (Muller)
Clinocardium ciliatum (Fabr.)
Hiatella arctica (Linn.)

Macoma brota (Dall)
Macoma calcarea (Gmelin)
Macoma nasuta (Conrad)
Nuculana sp.

Gastropoda:

Colus sp.
Lora sp.
Lunatia pallida (Broderip and Sowerby)
Neptunea lyrata (Gmelin)

Worm tubes
Balanus sp. (a barnacle)

The Adams Inlet and Muir Inlet samples come from a very large fauna that may amount to more than 200 species in a single bay. According to La Rocque the differences in the two assemblages are not particularly significant. Each assemblage is typical of the northern Pacific and they indicate no significant difference in climate from existing conditions. All of the species are shallow-water forms (2 to 18 m). There is no evidence of prolonged abrasion, and all forms appear to have been buried not far from their original habitat.

Other Organic Remains

Organic remains other than the faunal material are not common within the Forest Creek Formation. A spruce cone, dated at 11,170 years B.P. (I-2396), and other plant fragments were located in the upper part of the type section at Forest Creek, but so far these are the only plant materials found at the type section. At Granite Canyon, Section 66, 4 cm of peat overlie 0.5 m of sandy silt with pebbles at the top of the Forest Creek Formation. A sample (67-9) from this peat was analyzed for pollen to determine the climate at the time of formation of the peat.

The pollen spectrum (Table 7) from this sample, which is dated at 10,940 ± 155 years B.P. and is on top of the Forest Creek Formation, contains very little arboreal pollen. About 21 percent of the total pollen and spores are fern spores. According to Dr. P. Colinvaux who inspected the material, the spores and the pollen grains look fresh, and the fern spore concentration does not appear to be the result of differential weathering. The small amount of alder (*Alnus*) could have been transported several tens of kilometers (Colinvaux, 1967), and the pine (*Pinus*) may also have come from a great distance. Heusser (1960) found about the same percentage of pine in basal peat in the upper Montana Creek core about 80 km to the southeast near Juneau. The date on this material is 10,300 ± 400 years B.P. Although the species of pine was not identified in Sample 67-9, if the pine pollen were the lodgepole variety (*Pinus contorta*) it would not be surprising and may indicate the presence of this species in Glacier Bay shortly after deglaciation. Lodgepole pine (*Pinus contorta*) grows on poorly drained soil and is intolerant of shade. Heusser noted (1960, p. 48) that it has been reported in Glacier Bay by Cooper.

The dominant species in the spectrum are the ferns (Polypodiaceae) and horse tail (*Equisetum*), and these along with the sedges (Cyperaceae) and the grasses (Gramineae) constitute the bulk of the nonarboreal pollen and spores. These species suggest a low marshy area, probably without any trees for several tens of kilometers. After isostatic uplift exposed the Forest Creek marine clay, grasses, sedges, and ferns were established. The few grains of alders, trees of which may become established within 25 to 35 years in the present climate (Decker, in Goldthwait and others, 1966) may indicate that the deposit was formed either very soon after exposure or that the climate was too severe to permit growth of trees, possibly due to a nearby glacier.

Table 7. Pollen spectra from Adams Inlet, Alaska[1]

Pollen taxa	Percent of total pollen and spores (PS) and percent of pollen only (P) — Sample and radiocarbon age (B.P.)									
	67-9 10,940 ± 155		ST 12-5 --		58-13 1980 ± 100		58-12 1700 ± 100		66-5 1770 ± 100	
	PS	P	PS	P	PS	P	PS	P	PS	P
Alnus	3	5	29	33	5	7	4	10	16	29
Picea	-	-	48	55	43	69	34	75	30	54
Pinus	10	19	-	-	-	-	-	-	-	-
Ericaceae	6	11	+	3	-	-	+	+	+	+
Cyperaceae	14	25	+	3	5	8	+	+	6	10
Gramineae	11	20	+	+	+	+	4	9	+	+
Compositae	+	4	-	-	-	-	-	-	-	-
Polemonium	-	-	3	3	-	-	-	-	-	-
Polypodiaceae	21	-	13	-	15	-	54	-	41	-
Saxifragaceae	+	4	-	-	-	-	-	-	-	-
Lycopodium	+	-	+	-	23	-	+	-	3	-
Equisetum	23	-	-	-	-	-	-	-	-	-
Umbellifereae	3	6	-	-	-	-	-	-	-	-
Labiatae (?)	+	+	-	-	-	-	-	-	-	-
Other	3	6	+	+	8	14	+	3	+	4
Pollen and spores counted	188		307		200		349		300	

[1]Peat and forest duff samples collected and analyzed by the writer. Sample 67-9 is from peat on Forest Creek Formation (marine clay); sample ST 12-5 rests on 21 cm of weathered Granite Canyon till; samples 58-13 and 58-12 are from the forest beds near and at the top, respectively, of the Van Horn Formation; sample 66-5 is from a forest layer beneath the Adams Formation. *Senecio* and *Artemisia* comprise the Compositae in the spectra. Plus sign indicates 2 percent or less; minus sign indicates not detected.

A bed of ash occurs in sections 67C and 67D (Fig. 2); at these places the ash is 0.3 m and 0.5 m thick respectively. Section 67D, on the south side of Granite River, is better exposed although the lower 2.5 m above the river are slumped. In this section the ash layer is 1.3 m below gray clay and silt of the Forest Creek Formation, which is topped by 2 m of silt and stratified brown sand (dip S 6° W at 10°) that grades laterally into deltaic sands. These beds contain channel fillings from the lowest of several overlying coarse gravel units totaling about 5 m in thickness. Beneath the ash 0.5 m of gray silt and clay is above the slump.

The ash in Section 67D (Fig. 13) is laminated with a gray fresh surface showing some black organic mottles, and grayish-white to light brown on the oxidized surface. Worm burrows 3 mm to 4 mm in diameter contain a green material that is finer grained than the enclosing ash. The ash, of which glass constitutes the major part, appears void of macrofossils. The upper 0.2 m of ash is darker and may be mixed with other silt. A 5-mm-thick black layer at the base of the ash contains more organic material than the other parts of the ash; however, a 150-g sample submitted for radiocarbon dating yielded only 1/100 of the required organic material.

Grain Size Distribution

Sample 67C-4 was analyzed using both pipette and hydrometer methods. A specific gravity of 2.44 g per cc, determined with a pycnometer, was used in the calculations. The weight percentage of ash in the various fractions of the sand-clay range are given in Table 8.

Table 8. Mechanical composition of the ash (weight %)

Sand (mm)					Silt (mm)				Clay (μ)		
				Total				Total			Total
1.0-0.5	0.5-0.25	0.25-0.10	0.10-0.05	1.0-0.05	0.05-0.02	0.02-0.005	0.005-0.002	0.05-0.002	2-.2	<.2	< 2
Pipette method											
0.1	0.0	0.2	4.9	5.2	16.0	51.6	18.1	85.7	8.1	1.0	9.1
Hydrometer method											
0.0	0.0	0.0	3.0	3.0	22.0	52.0	16.0	90.0	--	--	7.0

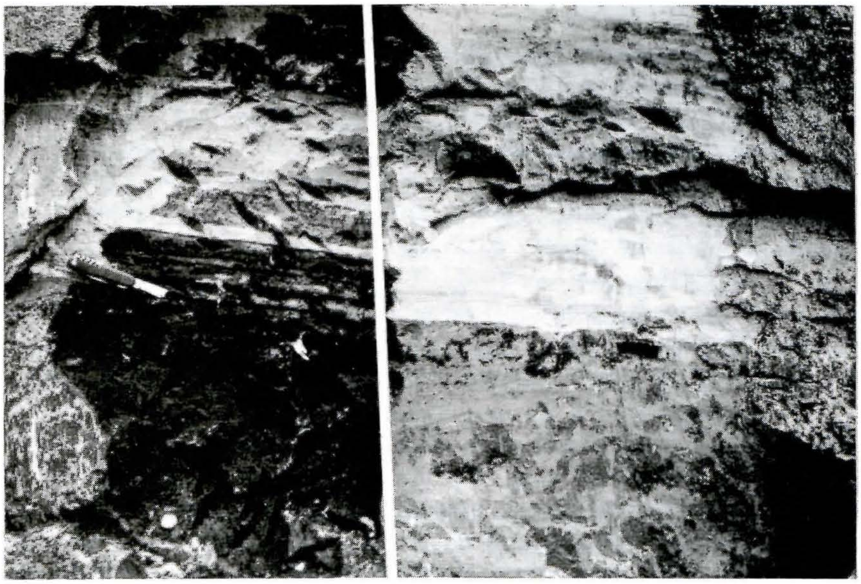


Figure 13. Volcanic ash unit within the Forest Creek Formation at Section 67D. Right side of photograph shows the weathered surface; the left side is unoxidized. Note fossils in the marine silt and clay below the ash.

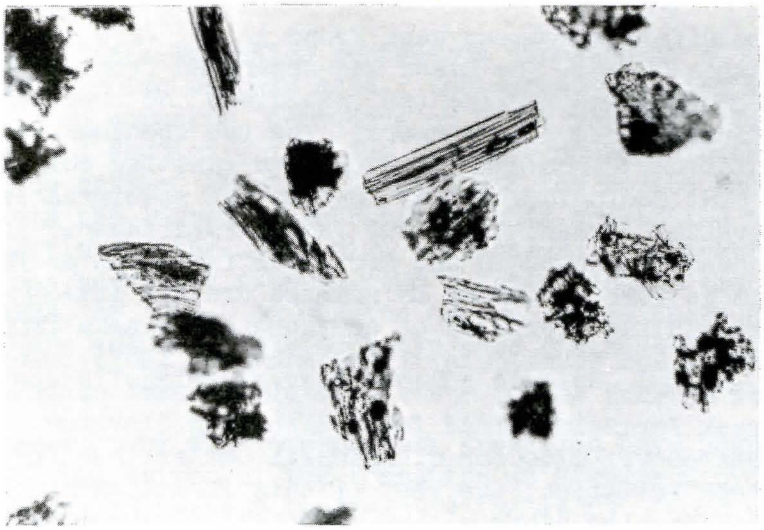


Figure 14. Volcanic glass fragments from the Forest Creek volcanic ash unit. Magnification about 125X.

Characteristics of the Glass

Glass shards are generally clear with no visible signs of alteration. Vesicular shards and fine threads of glass are common (Fig. 14). The index of refraction of the glass is approximately 1.51, which indicates an acid rock with about 68 percent silica (George, 1924). The corresponding specific gravity of such a glass is 2.39 g/cc (George, 1924), which is, as expected, less than the measured value of 2.44 g/cc for the whole ash including the heavier phenocrysts. X-ray diffractograms of ground ash were made using oriented specimens on ceramic plates and a random sample in a rotating sample holder. Feldspar and quartz peaks were identified, but no clay mineral peaks were present. Results of X-ray spectrographic analyses of selected elements are given in Table 4.

Mineralogy

Phenocrysts in the volcanic ash make up about 10 percent of the +250 mesh fraction. Of the heavy minerals (sp. gr. > 2.96) identified, pyroxenes and amphiboles were equally abundant and together constituted about half of the sample. Almost all of the orthopyroxenes and many of the other grains were coated with glass. Biotite and glass shards with inclusions each comprised about 10 percent of the sample. Twenty percent of the grains were opaque. Several small cloudy amorphous grains may be intergrowths of feldspar and cristobalite. The light fraction of the ash, other than the glass, consists mainly of albite with a minor amount of potassium feldspar.

Microfossils

Two Foraminifera specimens were found in the -250 mesh (< 0.061 mm) glass from Sample 67D-5. The glass was separated from the ash using a mixture of bromoform and acetone (sp. gr. 2.40) and treated with concentrated hydrofluoric acid for 24 hours. The two specimens are probably juvenile elphidiids. According to Ruth Todd (written communication, 1968) of the U.S. Geological Survey, one of the specimens is probably *Elphidium clavatum* Cushman; the other is unidentifiable. This species is well known and it occurs in the Arctic as far south as British Columbia today. It is rare to common in samples dredged (21-393 m) from the Gulf of Alaska and the fiordland of southeastern Alaska from bottoms of sand, silty clay, and pebbly clay (Todd and Low, 1967). In marine sediments elevated as much as 150 m and probably in part correlative with the Forest Creek Formation, Todd found *Elphidium clavatum* dominant. Although no search was made for microfossils other than in the ash of the Forest Creek Formation, the rest of this formation may be expected to contain this species along with some of the eleven other species (Todd and Low, 1967, p. 9) that comprise most of the fauna of the elevated marine deposits.

Age and Possible Source

Peat on top of the Forest Creek Formation at Section 67, which is approximately 200 m upstream from the sections containing the volcanic ash and at nearly the same elevation, has a radiocarbon date of 10,940 ± 155 years B.P. (I-2395). A spruce cone from the Forest Creek Formation at Forest Creek, where no ash has been found, is 11,170 ± 225 years B.P. (I-2396). These dates suggest an age of approximately 11,000 years B.P. for the deposition of the ash. Mt. Edgecumbe (not shown), 210 km (130 mi.) south of Granite Canyon on Kruzof Island, is the most probable source for the ash. The absence of eruptions from other volcanoes in the same general area, and the occurrence of what is believed to be Mt. Edgecumbe ash at Montana Creek near Juneau, 90 km to the southeast, suggest this origin. The eruption of Mt. Edgecumbe has been estimated at about 9,000 years ago based on a radiocarbon date of 10,300 ± 400 years B.P. (L-207D) on peat beneath the ash at Montana Creek (Heusser, 1960). Since this date is on material underlying the ash, the date of the eruption may actually have been about 10,300 years B.P. It is conceivable that Mt. Edgecumbe may have erupted as early as 11,000 years B.P. and continued to be active for a thousand years, or longer. Correlation of the Granite Canyon ash with Mt. Edgecumbe, using mineralogy and refractive index of the glass, may be possible upon completion of analyses on ash collected near Sitka (R. W. Lemke, written communication). A sample of the Granite Canyon ash is being analyzed by R. E. Wilcox of the U.S. Geological Survey along with Lemke's samples.

Origin and Age

The Forest Creek Formation represents deposits of an invading sea following deglaciation of this area. Ice-rafted material may be present or absent in such deposits, depending on their proximity to the ice margin. At the type section all of the exposure contains pebbles; the visible difference between the till beneath and the marine sand and silt is the presence in the latter of unbroken mollusk shells and barnacle plates in growth position on the upper side of pebbles. The pebbles in the marine beds may have been dropped from floating ice in the form of an ice shelf, icebergs, or both. At Granite Canyon, where there are few pebbles in the clay, the ice front was probably some distance away and iceberg rafting was not effective in transporting pebbles.

Three sections of the Forest Creek Formation have been dated by radiocarbon. Wood on top of the formation at Granite Canyon gives a date of 10,940 ± 155 years B.P. (I-2395) (Table 6). Wood above the Forest Creek Formation at upper Forest Creek gives a younger date of 10,400 ± 260 years B.P. (I-1615). Three dates, two of them on shells, are available from the type section. One shell, dated 10,000 ± 220 years B.P. (I-1303) was collected 5 m above the base of the formation; the other, dated 13,960 ± 360 years B.P. (GXO-460), was collected 2 m lower than I-1303, but at an unknown distance above the base. A date on an allochthonous spruce cone collected in the Forest Creek Formation 0.6 m below its top is 11,170 ± 225 years B.P. (I-2396), which provides a maximum age for at least part of the Forest Creek Formation. Considering this date and the two minimum dates on autochthonous material at two other sections, a good estimate of the age of the formation is 11,000 years B.P. Besides the experimental error associated with the dates, and the possible difference in time lag before establishment of organisms, differences in the dates may also reflect the diachronous nature of the Forest Creek Formation.

Deposits That May Correlate With Forest Creek Formation

In the upper regions of Glacier Bay marine silt and clay of the Forest Creek Formation is present in five exposures ranging in elevation from 3 to 59 m. Other deposits of marine silt and clay with shells, at the edge of the mountains east of Gustavus, were found by D. J. Miller at elevations from 6 to 60 m (Brew, oral communication). On Douglas Island the maximum elevation of marine deposits is 150 m (Twenhofel, 1952), and at Montana Creek, 24 km northwest of Juneau, the maximum elevation is 120 m (Heusser, 1960). Marine silt and clay with pebbles and shells, in appearance very much like the Forest Creek Formation, crops out 200 m from the end of Montana Creek road at an elevation of approximately 135 m. Although these marine deposits indicate a high stand of the sea following the retreat of Wisconsin ice, there are several reasons why they might not correlate either in time or elevation. The ages for these deposits might not agree because of different times of deglaciation. A lag in establishment of the organism used for dating and experimental errors may preclude correlation of some isochronous unit. Differences in elevation may be due to different rates of uplift caused by isostatic and tectonic forces in these different areas. Another problem in correlating marine deposits other than shoreline materials is the unknown depth at which the deposits were formed. The experimental errors in dating could result in different elevations with the same radiocarbon age. For example, a 225-year error in a radiocarbon age determination coupled with an uplift of 4.0 m per century (Hicks and Shofnos, 1965) or 4.5 m per century (Goldthwait and others, 1966) could result in a 9 to 10 m difference in elevation. Considering these facts, it is not unexpected that there are different elevations and different dates for the Forest Creek Formation and other marine deposits in this part of Alaska. Even so it may be inferred that at least the highest deposits in a particular area were formed soon after deglaciation of that region, and thus should be considered as belonging to the same formation.

In Glacier Bay three separate occurrences of Forest Creek sediments have an elevation close to 30 m. If at the time of deposition sea level was 37 m lower than today (Shepard, 1963), this indicates a total rise of land in this part of Glacier Bay of at least 67 m.

Lower Member of the Van Horn Formation

The lower member of the Van Horn Formation at Van Horn Ridge on the east side of Muir Inlet was defined by Haselton (1966) as a poorly sorted and poorly bedded yellowish-brown gravel. Deposition of this unit was suggested to have been intermittent, as indicated by the presence of two forest beds underlain by sands and silts with incipient soil horizons. Bedding is essentially horizontal and cut-and-fill structures indicate braided streams. Most of the pebbles in this unit are weathered; the percentage of totally weathered pebbles is even greater adjacent to organic layers (Haselton, 1966). These gravels began forming during early Hypsithermal time, and in Muir Inlet are bracketed by dates of 7,075 ± 250 years B.P. (I-84) and 3,290 ± 55 years B.P. (Y-303). The lower member of the Van Horn Formation is present in sections in Muir Inlet from Canyon Creek, 3 km north of Nunatak Cove, to Morse River. South of Morse River the gravels are lithologically different and may have a source area west of Muir Inlet. A. T. Ovenshine believes that the reddish-brown gravels at U.S. Coast and Geodetic Survey Station "Bull" are the same age as the middle Van Horn member (Haselton, 1967), and a date of 2390 ± 110 (I-3398) on allochthonous wood beneath clays near the base of a section at Station "Quill" on the west side of Muir Inlet supports this hypothesis.

The lowest gravels in Adams Inlet are correlated with the lower member of the Van Horn Formation of Muir Inlet on the basis of similar lithology--nearly horizontally bedded, fine grained, poorly sorted, dark brown color; in many places the pebbles are highly weathered as in Muir Inlet. Pebble counts in the Adams Inlet and tributary valleys show influence of material from these valleys and thus are variable as is the case in Muir Inlet.

Based on radiocarbon dates of wood in the Van Horn Formation in nearby Wachusett Inlet, Goldthwait (1963) determined the rate of building of the sand and gravel to have been between 1 and 5 cm per year. These rates, the first dated by radiocarbon, may also be applicable to the Van Horn Formation in Adams Inlet.

Distribution and Nature of Deposits in Adams Inlet

In Adams Inlet the gravel of the lower member of the Van Horn Formation is medium to fine grained, poorly sorted, and poorly stratified. In some sections coarse sand beds are common, and Section 58 contains three units of rhythmically bedded clays and silts as much as 3 m thick. A sieve analysis on Sample 1A-1 from this unit in White Valley (Fig. 15)

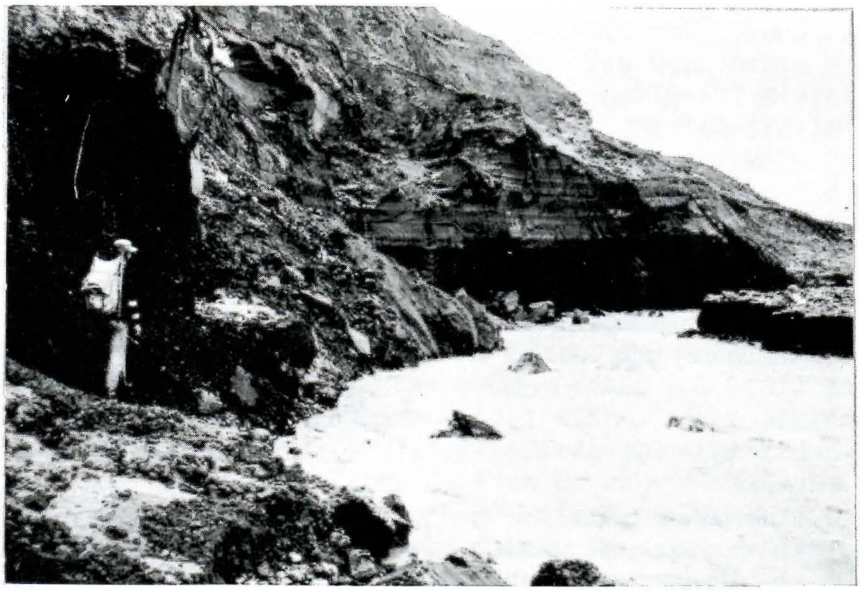


Figure 15. Van Horn Formation, lower member, at Section 1A in White Valley. Gravel is overlain by rhythmic silts and clays of Adams Formation.

gives a mean particle-size diameter of -2.25 phi (4.7 cm) and a phi deviation measure of 2.25 phi, illustrating the fine nature of the gravel and the poor sorting, respectively (Inman, 1952). Other parameters for the same sample are: phi median diameter, -3.05 phi; phi skewness measure, -2.36; second phi skewness measure, -2.06, and phi kurtosis measure, 0.422. The color of the gravel varies with location relative to the valleys from which the pebbles were derived, but it is generally dark brown to reddish brown. The lithologies of 23 pebble counts reflect source areas south and southeast of the Inlet. Graywacke, volcanic rocks, hornfels, and dike rocks comprise the assemblages with the exclusion of plutonic igneous rocks except for a few samples from the upper part of the unit, particularly on Adams island.

The lower member of the Van Horn Formation is present throughout Adams Inlet, but is best exposed in several sections on the south side. At Section 58-A as much as 20 m of this unit are exposed, with forest beds at the top and 1 m below the top. Logs are also present in this member approximately 3 m above high tide. The base of this unit is not exposed in most sections but in the upper part of Adams Valley near Section 75 local gravels of the lower member of the Van Horn, with stumps at the base, overlie bedrock, and at Section 67 weathered gravels thought to be Van Horn overlie the Forest Creek Formation. In the central part of Adams Inlet the lower member of the Van Horn Formation is overlain by lacustrine sediments and till of the Adams Formation. In several valleys this lower unit rises up-valley at about the present gradient of the streams, and in White and Pyramid valleys deltaic foreset bedding is present at elevations of 168 m to 185 m, and 200 to 210 m, respectively. The middle and upper members of the Van Horn as identified in Muir Inlet are not present in Adams Inlet.

Van Horn Diamicton

Another type of non-organic deposit in the Van Horn Formation is an 80-cm-thick diamicton that crops out only in Section 66. It consists of greenish-gray pebbly clay between two forest beds. The lower forest bed overlies 60 cm of fine cobbly gravel on weathered Granite Canyon till; the upper one is overlain by lacustrine deposits of the Adams Formation.

The most abundant clay minerals of the diamicton (Table 3) are illite and chlorite. The abundance of interstratified clays and quartz is the same as in the weathered Granite Canyon till.

Weathering and sorting may have occurred in the diamicton to produce the similarities in composition to the weathered Granite Canyon till; however, it is not known if the parent materials were the same and hence it is difficult to assess the degree of weathering in the diamicton. That weathering has occurred is evidenced by the low percentage of calcium found in the elemental analyses (Table 3). The Zr/Ca ratio is 0.096, lower than that for the weathered Granite Canyon till, but higher than that for the unweathered till. Changes in composition on leaching with HCl are also similar to those in the weathered Granite Canyon till sample; and there is very little decline in calcium on leaching. These facts indicate a weathered deposit. The diamicton appears in section to be homogeneous and to lack a well-developed soil profile, and thus may be a solifluction or colluvial deposit.

In support of a solifluction origin for the diamicton is the change in climate that occurred at about the time of formation of this deposit. The underlying peat, dated at 3850 years B.P. (Table 6) is of early Neoglacial age. It is inferred that increased precipitation or other climatic factors may have produced conditions favorable for mass movement of previously stable slopes. Although ice was advancing down Muir Inlet by this time (Haselton, 1966), it is unlikely that it had reached Adams Inlet. Even less likely is an advance down Granite Canyon by glaciers from the lower mountains to the east where precipitation is less than in the west.

Organic Deposits

Organic deposits, mainly in the form of well-developed forest beds with stumps as much as 60 cm in diameter in growth position, occur within, at the base (on bedrock), and at the top of the Van Horn Formation in Adams Inlet. Allochthonous stumps and logs also occur in the gravels. The forest beds contain forest duff or peat as much as 10 cm thick, and are underlain by leached sandy and silty gravel. Gravel advancing down the valleys buried and destroyed some of these forests growing on earlier deposits of river gravel (Fig. 16). The youngest forest was inundated by deposits of glacial Lake Adams (see section on Adams Formation).

Trees in the Van Horn forest beds in Adams Inlet were not identified; however, Van Horn forest beds of Muir Inlet have been studied by Cooper (1931) and by Goldthwait and Burns (Goldthwait, 1963). They recognized two kinds of Hypsithermal forests: a climax type that covered the valley walls, and a subclimax or unstable forest that covered the outwash plains. The climax forests were found to contain 29 percent *Picea*, 67 percent *Tsuga* and 4 percent *Populus*. The average composition of forests on former outwash plains was 58 percent *Picea*, 22 percent *Tsuga*, 19 percent *Populus*, and 1 percent *Alnus*; however, much variation was found. On the terminal moraine at Bartlett Cove, a 215-year-old subclimax forest was found to contain 9 *Picea* and 5 *Tsuga* per 10 m^2 (Goldthwait and others, 1966).

In an effort to determine climate and type of forests in Adams Inlet of 1700 to 1900 years ago, four samples of forest duff and peat, 4-10 cm thick, were analyzed for pollen. Two samples were from eastern Adams Inlet (58-12, 58-13) and one was from Granite Canyon (66-5). Another sample (ST 12-5) that was not dated but is probably similar in age to the others, was also analyzed. The results of the pollen analyses are given in Table 7, and the interpretation of the pollen spectra are given below. In general most of the forests were of the short-lived non-climax

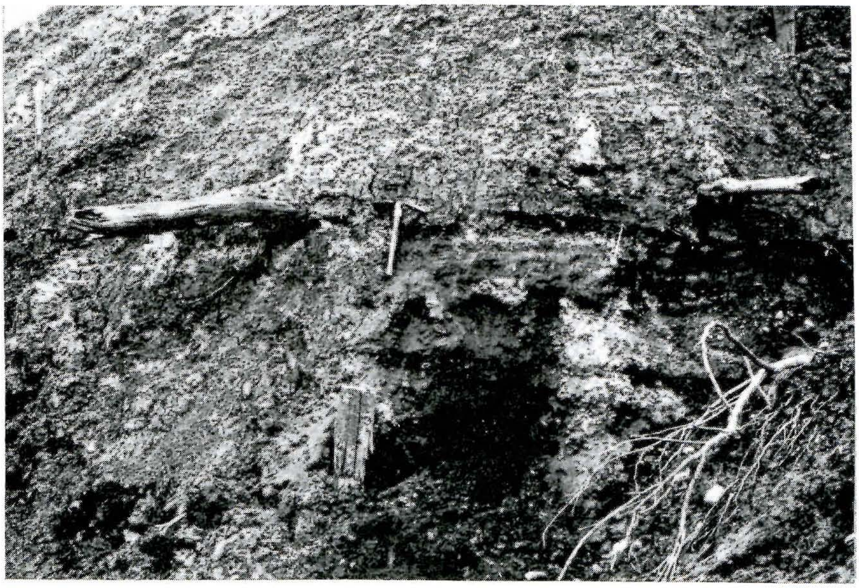


Figure 16. Two forest beds in the Van Horn Formation at Section 58. The stump from the lower forest bed is dated at 1,980 ± 100 years B.P. (I-2394). It rests on 2.5 m of lacustrine silt and clay and is overlain by 1.3 m of brown weathered gravel and silt. The outside of the 60-cm-diameter stump from the upper forest bed is dated at 1,700 ± 100 years B.P. (I-2687). This unit is overlain by lacustrine beds of the Adams Formation.

types, as would be expected from the field occurrences.

Interpretation of Pollen Spectra

Sample 58-13 is from the lower forest bed, in the upper part of the Van Horn Formation (Fig. 16). The pollen spectrum (Table 7) suggests a spruce forest in the area. The forest, which was established on Hypsithermal gravels, was destroyed by burial with gravel of the Van Horn Formation that continued to advance down the valleys. Absence of hemlock Tsuga heterophylla), which is present in the 215-year-old forest at Bartlett Cove and also in Hypsithermal forests in other parts of Glacier Bay (Goldthwait, 1963), suggests that the forest was of short duration. The largest stump contained only approximately 110 growth rings, which also supports the suggestion of a short-lived forest.

Sample 58-12 is from a forest bed about 1 m above the one from which Sample 58-13 was taken and is younger by about 280 radiocarbon years (Table 6). The interpretation here is similar to that for the lower forest bed--a spruce forest of short duration that did not reach the climax stage of development. The forest layer, which contains stumps as much as 60 cm in diameter (with approximately 180 growth rings) was inundated by the filling of glacial Lake Adams.

The relatively low percentage of spruce pollen in Sample 58-12 may be attributed to the correspondingly higher number of fern spores (polypodiaceae). Discounting some of these fern spores increases the relative percentage of spruce pollen to a level comparable to that in the lower forest layer. The high number of fern spores may be due to a residual concentration effect (Colinvaux, personal communication).

A forest layer overlain by the Adams Formation, resting on the Van Horn diamicton, and containing stumps as much as 40 cm in diameter (Sample 66-5) was dated at 1700 years B.P. The pollen spectrum is very similar to that of the nearly contemporaneous Sample 58-12 with spruce the second most abundant species after the sedges. The same interpretation, that of a spruce forest, is applicable. The higher percentage of alder in this sample compared to most of the other forest beds is probably not significant.

An undated forest bed on the west side of Goddess river near Section 42 is the source of Sample ST 12-5. This forest bed rests on 21 cm of weathered Granite Canyon till, and is overlain by the Adams Formation. Spruce and alder dominate the pollen spectrum and together total almost 80 percent (Table 7). The data indicate that a spruce forest was established in this area and alders may have been growing within about 10 km. The low percentage of spores also supports this interpretation. Logs 20 cm in diameter occur in the forest layer. The occurrence of Polemonium, a tundra element, is unusual in this situation. It may mean a local open area. Heusser (1960, p. 92, 118) has noted this genus in several scattered locations along the coast, and in his Munday Creek profile

(Heusser, 1960, Fig. 16) it is found in the early Postglacial (9000 B.P.) associated with alders, ferns, and sedges.

Age and Origin

The forest bed at the top of the lower member of the Van Horn Formation in Section 58 is dated at 1700 ± 100 years B.P. (I-2687), and another forest bed 1 m lower in the same section is dated at 1980 ± 100 (I-2394). Infilling of valleys south and east of Adams Inlet continued after the beginning of Lake Adams (described later) and formed deltaic deposits in several of these valleys. Thus, deposition of gravels of the Van Horn Formation persisted for at least several hundred years beyond 1700 B.P. The oldest date on wood within the Van Horn Formation in Adams Inlet is 3710 ± 110 years B.P. (I-164)(collected by D. J. Miller and submitted by R. P. Goldthwait) at 12 m below the top of the formation near Section 58. The oldest part of the Van Horn Formation in Muir Inlet is 7075 ± 250 years B.P. (I-84) at Westdahl Point and the youngest date is 3290 ± 55 years B.P. (Y-303) at Hunter Cove. This period from about 7100 B.P. to 4200 B.P. (ten dates averaged for end date) has been assigned to the Hypsithermal (Goldthwait, 1966) and represents the time of deposition of most of the lower Van Horn Formation in Muir Inlet, at an average rate of 1.4 cm per year (Goldthwait, 1963). Evidently only the latter part of the post-glacial filling is exposed in Adams Inlet, the older deposits of gravel being deeper in the section beneath Adams Inlet. What happened in Adams Inlet between 11,000 B.P. and 3700 B.P.? Granite Canyon is the only place where a clue to the events of that time is given. At Section 67 a 3-m-thick unit of coarse gravel overlies peat dated at 10,940 ± 155 years B.P.; at Section 66, 1 km upstream the weathered Granite Canyon till is overlain by 0.6 m of gravel with a forest bed at the top dated at 3850 ± 110 years B.P. (I-3068). Apparently gravel was deposited in the lower parts of the Inlet after recession of the sea during Forest Creek time but in the vicinity of the Inlet never reached as high as 75 m above present sea level. This is suggested by the low elevations of this formation in Section 67, and by the weathered Granite Canyon till covered by a thin layer of gravel, probably talus in part, at an elevation of 75 m in Section 66.

Adams Formation

Definition

Unit 4 of the composite section consists of massive to rhythmically-bedded lacustrine clays and sandy silts in which are interspersed several units of clayey till, and also some fine sand units toward the top. This lacustrine-till complex is here called the Adams Formation, and Section 58 is the type section. The top of this section, at an elevation of

199 m (Fig. 2, 16), is 1.9 km N 30° E of the top of Tree Mountain. The coordinates of this section are 58° 53' N, 135° 47' 54" W.

The Adams Formation is set up as a separate rock-stratigraphic unit because it has distinctive lithologic characteristics (rhythmic silts and clays with clayey till, pebble composition, and elemental composition) that make the unit mappable. A small part of the formation consists of till that was deposited by Neoglacial ice and thus by some writers might be considered to be part of the Glacier Bay Formation, and by others it might be considered to be part of the Van Horn Formation. Although the close relationship is recognized, the distinctiveness of this lacustrine-till complex, the extent of the unit, and the importance of the indicated fluctuations of glacier ice are considered important enough to designate this practicable unit as a formation.

The glacial lake in which this formation was deposited is here named Lake Adams. This formation, of which lacustrine deposits make up more than 70 percent, exhibits steep slopes and cliffs in section. Because the formation is relatively impermeable, many springs have their source at the top of this unit. A brown iron stain covers some of the cliffs.

Beneath the Adams Formation is the lower member of the Van Horn Formation (Fig. 17). The contact is generally very sharp. Laminated silts and clays generally overlie gravels or forest beds, stumps of which occasionally protrude as much as 2.5 m into the Adams Formation. A thin layer (5 cm) of till forms the base in parts of upper Adams Valley.

The contact with the overlying Berg Formation is gradational through medium-fine sands, some of which contain gravel pockets and disturbed bedding. The top of the uppermost clay, silt or very fine sand is taken as the top of the Adams Formation.

Distribution

The Adams Formation is widespread in the lowland areas of the map area, but thicker and better exposed south of Adams Inlet (Plate I). In this part of the map area it is as much as 66 m thick, and it occurs from lower White Valley on the west to Pyramid Valley on the east, and as far south as Endicott Gap. Exposures are particularly good on the east side of Goddess river. The formation also occurs in Berg Valley, Granite Canyon, and the banks of lower Seal River. It was not located on the north shore of the Inlet except north of the peninsula in the west end of the Inlet. The elevations of this unit vary from about 5 m below high water, as at Section 11, to 165 m above tide near Endicott Gap.

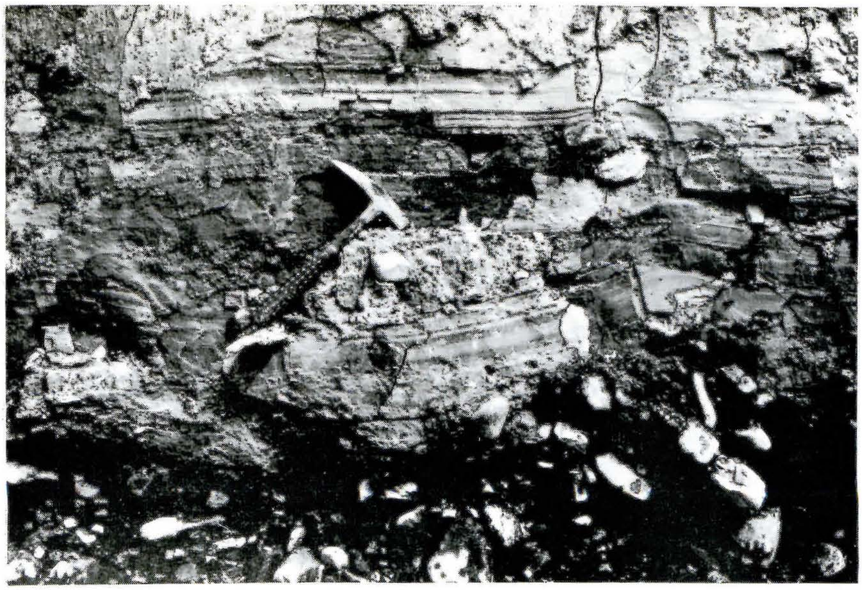


Figure 17. Silts and clays, with pebbles, of the Adams Formation overlie gravels of the Van Horn Formation near Section 5. Hammer rests on block of till in rhythmites.

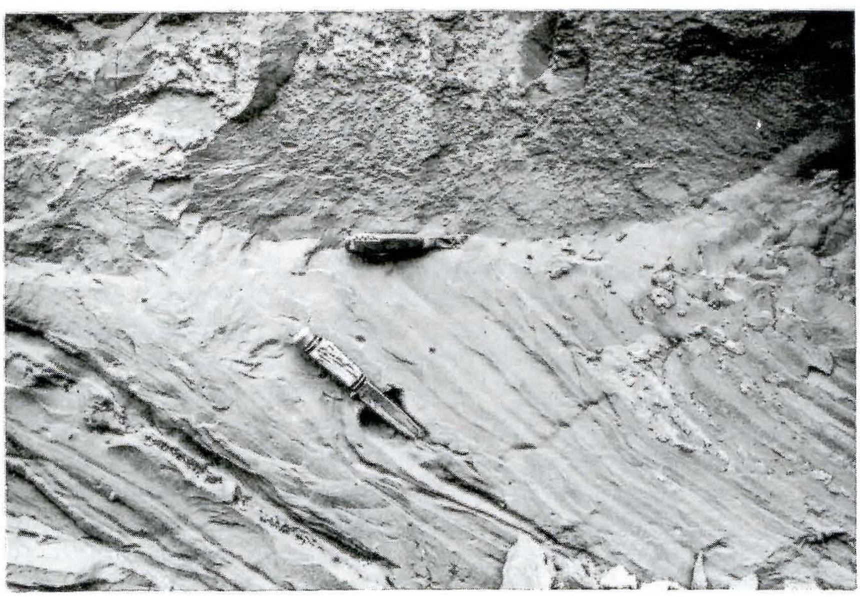


Figure 18. Clayey till over deformed clays and silts in the Adams Formation near Section 5. Glacier flow from left to right up Adams Valley.

Nature of Deposits

Most of the formation is comprised of rhythmically-bedded silt and clay. Each rhythmite, that is, a lower silt layer and an upper clayey layer, may represent an annual layer; if so, they could be called varves. The rhythmite deposits average about 1-5 cm thick but some are as much as 70 cm thick. In some places, as in lower Goddess valley, there are several meters of massive silty clay.

Most of the clays of the Adams Formation are gray or greenish-gray; however, in the lower meter of the formation in White Valley, on Adams Island, and near Section 58 some clay and silt layers are dark red. This red color is similar to that of the fine-grained portion of the underlying Van Horn gravel and represents continued inwash of red sediments during the early stages of glacial Lake Adams. In many parts of the formation, particularly that part in Adams Valley, there are ice-rafted pebbles and cobbles, often striated, and blocks of till (Fig. 17). Much of the bedding is disturbed and wraps around large boulders. Buff silt stringers, layers of concentrated fine-grained organic material, fragments of wood, and logs also occur in many parts of the formation. Peaty organic material is particularly common in sections on the west end of Adams island. Penecontemporaneous folding of beds also occurs. In Granite Canyon (Section 64) the amplitudes of such folds reaches 60 cm. Folding has also been caused by overriding ice as in several places in upper Adams Valley.

Near Section 5 laminated sediments are folded and tilted, and indicate pressure from the north. In some cases till, or hard, indurated reworked clay, has been left in contact with the folded sediments (Fig. 18).

Several hundred meters upstream from Section 4 on the east side of Adams River, a large thrust block, 2 m by 8 m, of poorly sorted medium-grained gravel is surrounded by clayey till (Fig. 19). Bedding is poorly defined in the gravel, but some indication of layering is visible in the wet silty layers in the lower block. Bedding in this block dips upstream, reverse of the normal horizontal to slightly downstream dip of the Van Horn Formation in Adams Valley. Tilting of the gravel blocks was the result of deformation by the ice. A glacier flowing from Adams Inlet advanced up glacial Lake Adams in Adams Valley, probably as floating ice, touched down, if it were not already riding along the bottom, and began to push up clay in front of it. With continued erosion and pressure, folding in the underlying gravels was initiated and a large block of the Van Horn Formation began to lift up and push into the till (Fig. 19, right side of photo). Finally the force was too great for the unconsolidated material (probably not frozen), and part of the underlying gravel sheared off and overrode the upfolded portion. A trail of gravel particles and a small (20 cm) block of gravel were left in the clayey till in the lee of the advancing gravel block. Clayey till with some remnant

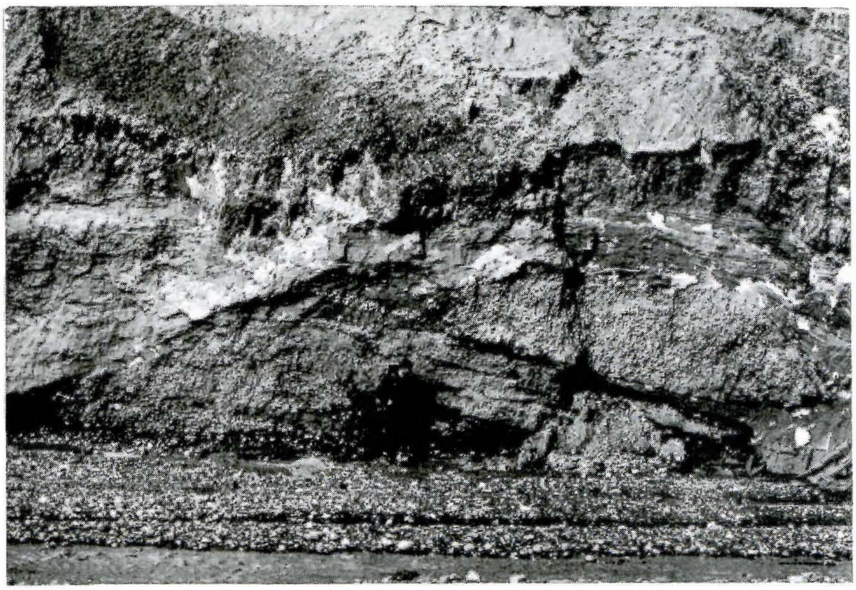
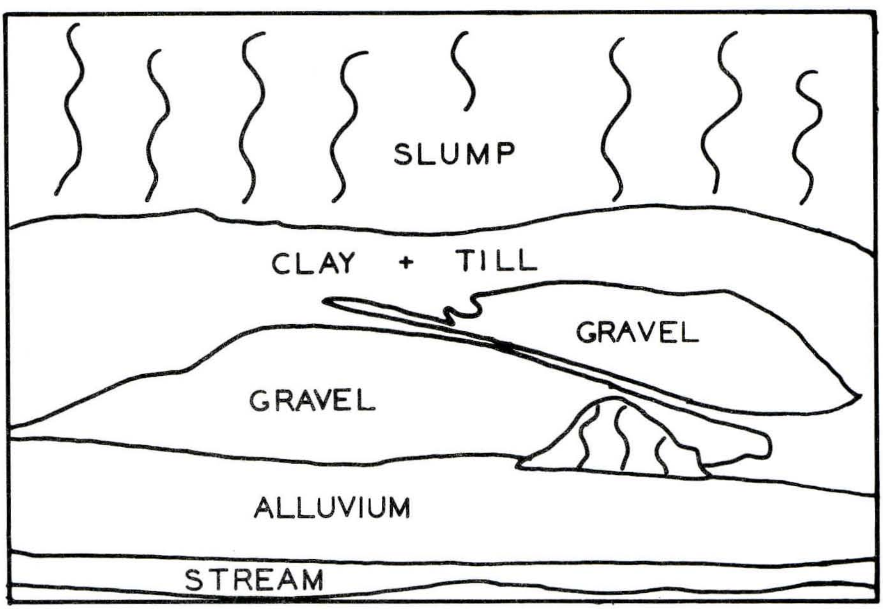


Figure 19. Deformed clays and gravels on the east side of Adams Valley. Gravels of the Van Horn Formation, which are normally less than a meter above river level in this area, have been deformed, bulged up and sheared off by ice moving from the north (left) to Adams Valley. Materials are indicated by the line drawing below.



bedding remains on both sides of the gravel. Laminated lacustrine beds overlie the till and gravel. Presumably these were deposited after a retreat of the ice northward beyond this point, or after the glacier resumed floating here.

Grain Size Distribution

Four samples from the fine fractions and one from the coarse fraction of rhythmites showed an average of 56.2 percent silt and 43.8 percent clay (Table 1). No sand was recorded, even in the sample of coarse material, which had 68 percent silt, and all samples plot along the side of the triangular coordinate diagram for grain-size distribution (Fig. 8). Because the clayey laminae make up a small percentage of the Adams glaciolacustrine material, these averages are not really representative of this unit of the Adams Formation. Thus biased sampling is one reason why grain-size analyses indicate that this unit is more clayey than the Adams till (Fig. 8) and the Forest Creek Formation.

Pebble Lithologies

One pebble count was made on pebbles in this glaciolacustrine unit from south of Adams Inlet at Section 48. In composition it is close to the average for Adams till in Adams Inlet but has more graywacke and fewer plutonic igneous pebbles. The difference is probably due to dilution from local graywacke bedrock (Fig. 5) and as a result the PI/M ratio for this sample is only 0.47.

Clay Minerals

Illite and chlorite are the major constituents of the clay minerals in the Adams lacustrine unit and are present in equal amounts (Table 3). Expandable clays and quartz each make up one part in ten. In this unweathered sample all clay mineral components are considered to be of detrital origin. The clay mineral composition is not very different from that of the Forest Creek Formation. The lacustrine clays are also similar to the clays of the Adams till as would be expected of material in the same formation. The till has more montmorillonite than the lacustrine clay, but this difference is not deemed significant.

Elemental and Carbonate Analyses

Elemental analyses (Table 4) of the lacustrine material show it to be not too different in composition from other units examined, although the percentage of Ti and K are higher than younger units and lower than older units. The unit is characterized by the highest percentage of Zr of the units investigated. All of the units in this study have Zr values two to three times higher than those found in several studies of tills and soils in the midwestern United States (Wilding, personal communication). The different bedrock in Adams Inlet may explain this difference. With HCl leaching, the relative amounts of all elements decrease, except

Ti and K. This is the reaction found in other units except the weathered Granite Canyon till and the Glacier Bay till, both of which showed relative increases in Zr. The wide variation in composition of the silt and sand fraction between Adams lacustrine material and Adams till, both with approximately the same clay mineral composition, may be due to greater abundance of sand size minerals in the till.

Three samples of lacustrine silt and clay averaged 10.4 percent total carbonate (Table 5). This is about the same as the total carbonate percentage for the Adams till (9.9 percent). The average total carbonate content and the calcite-dolomite ratio of the lacustrine unit is little different from the glaciomarine unit.

Till Units

Nature of Deposits

Till comprises less than 30 percent of the Adams Formation and occurs as distinct beds as much as 6 m thick and traceable, at least in Adams Valley, for 500 m. Till is also present as discontinuous units ranging from 2 to 0.05 m in thickness. Some of the blocks of till occur in laminated clays (Fig. 17). The till is mostly light to dark gray.

Grain-Size Distribution

The till contains very few pebbles and thus is difficult to recognize as a till. Mechanical analyses (Table 1) of ten samples from Adams Inlet average 18.5 percent sand, 66.2 percent silt, and 15.3 percent clay; all but one of these samples falls within the silty loam classification (Fig. 8). This till is distinguishable from other tills in the area and from the lacustrine units of the formation on the basis of the grain-size distribution in the < 2 mm fraction (see section on Glacier Bay Formation).

Pebble Lithologies

The pebbles collected from exposures of till of the Adams Formation in Adams Inlet are similar to those of the Glacier Bay till, but more are plutonic igneous rocks and there are slightly fewer limestones and dike rocks (Table 2). This similarity would be expected if both tills were derived from bedrock north of the Inlet. These two tills can, however, be differentiated in Adams Inlet on the basis of their ratios of plutonic igneous rocks to metasedimentary and sedimentary rocks (PI/M ratios). The Adams till has a ratio of 0.69 and the Glacier Bay till has a ratio of 0.44 for samples from Adams Inlet only. The range of ratios for these tills in other valleys is quite large (Table 2). Much variability in the composition of individual and average pebble counts and in the range of PI/M ratios (Table 2) for these tills exists in valleys tributary to Adams Inlet. Although the number of samples is limited, it appears as if the PI/M ratio cannot be used to distinguish Adams and Glacier Bay tills in the area surrounding Adams Inlet.

The Adams till shows the highest average PI/M ratio of the three tills (Table 2). This is unexpected as this formation overlies the Lower Van Horn gravels, which in Adams Inlet and valleys to the south has a PI/M ratio that ranges from 0.76 to 0.00. The high PI/M ratio (up to 1.2) in Adams Valley, where graywacke is present on both sides, can be explained when the glaciolacustrine nature of the Adams Formation is considered. The glacier that deposited the Adams till came from the north (high plutonic igneous percentages) and entered the lake in Adams Inlet. At times it was a floating ice tongue and thus did not incorporate, except from north of Adams Inlet, any metasedimentary rocks. At times the glacier grounded, but the clay deposits on the bottom of the lake were thick enough to prevent incorporation of the underlying Van Horn gravel, and thus these gravels did not supply metasedimentary material to the till. In only one place were the gravels seen to be disturbed by ice movement (Fig. 19).

Clay Minerals and Elemental Analyses

Clay minerals of the till (Table 3) are essentially the same as those of the glaciolacustrine material. This similarity is not unexpected since the lacustrine material was probably derived from the same englacial material as the till and some of the fine fraction of the till may have been incorporated from the lacustrine clay. Conversely, since these samples are from the same formation in the same section, separated by a vertical distance of 20 m, the similarity of composition serves to confirm the vertical homogeneity and interrelation of members of this formation.

The Adams till has less quartz and illite, and more chlorite and expandable clays, than the other two tills. Many more samples would have to be analyzed to determine if the clay mineral differences between till units were significant and constant over a wide geographic area, in order to use clay minerals for differentiation of till sheets. Horizontal variation in composition within the till sheet was not determined.

The elements in sand and silt fraction of the till are similar to those of the lacustrine units; differences as already mentioned probably are due to the higher percentage of sand in the till. The Adams till has a lower Zr percentage than either the Glacier Bay or Granite Canyon tills; Ti percentage for Adams till is between the other two tills.

Glacier Movements

Till Fabric

Fabric analyses were made at 9 localities in till units of the Adams Formation. These were mainly in sections on the south shore of

Adams Inlet but several were made in Adams Valley. The resultant directions are given in Fig. 10, and the fabric diagrams are given in Fig. 9.

Fabrics in the Adams till were generally not as strong as those in the other two tills in the map area, and in one case (67-12) no single maximum was present. The reason for this lack of a strong fabric is not known. It may be due to something inherent in a deposit formed by a glacier which is alternately floating and sliding over clays. The small median grain size of the till and lack of pebbles may be causes of this poorly defined maximum. Changes in ice flow direction may have resulted in the partial destruction of one maximum and the initiation of another. Incomplete readjustment of pebbles to the forces of this different ice flow and an ablation source could result in the wide girdles seen in some of the fabrics (Fig. 9). Changes in flow direction may have been common in a glacier moving into a lake, occasionally grounding, and advancing and retreating up and down valleys. Competition from a lobe at the entrance to Adams Inlet may also have caused changes in ice directions in Adams Inlet.

The inferred ice-flow direction at the entrance to White Valley (Fig. 10) is from east-northeast (Fig. 9, 67-03AT) and northeast (67-08AT). The maxima in these fabrics are broad and cover about 90° of arc. The fabric (67-11AT) south of Camp Adams is also of this type and shows about the same trend. Approximately 900 m east of 67-11AT, in a small gully also on the south side of the Inlet, the fabric 67-12AT was obtained. Ice flow appears to have been from the west and north. It may represent a westerly flow direction with a remnant northerly component.

The strongest fabric is 67-13AT, which is north of Tree Mountain. Flow from the west is indicated. An east-west flow would be expected here against the side of the mountain. Fabric 67-15AT was done at the entrance to Adams Valley (Fig. 10). The maximum is parallel to Adams Valley but no preferred dip is present. A similar situation exists for 67-16AT on the northeast corner of Tree Mountain.

In Adams Valley, 2 km from the Inlet, two fabrics were done in the Adams Formation. These fabrics (67-21AT and 67-21A-AT), give broad girdles over 90°. Slight northerly dips are indicated. A micro-fabric, Section 49A-1 (Fig. 27), has a north dip and indicates ice flow out of Adams Inlet toward Endicott Gap. Ice-flow directions in Adams and Goddess valleys were similar during deposition of the Glacier Bay Formation.

Age and Origin

As indicated by a date on a stump protruding from the underlying forest bed 2.5 m into the base of the Adams Formation, deposition of lacustrine materials began 20 m above present sea level about 1700 years ago. No radiometric date for the base of the overlying Berg Formation is available; however an estimate of the minimum time for deposition of the Adams Formation can be made by considering the rhythmic silts and clays as varves. Each pair of laminae may represent the deposit of one year. At Section 58 (Fig. 2), there are, among other things, 30 m of rhythmites 2-5 cm thick. Thus the range of time for deposition of this part of the Adams Formation would be 600-1500 years. This does not include, however, the 21 m of fine to medium sand, some of it rhythmically bedded, deposited in two units within and at the top of the formation. Also not included in the time estimate is a till unit, 17 m thick, in the middle of the formation.

The unusual nature of the Adams Formation, a lacustrine-till complex, indicates that there was an ice-dammed lake in Adams Inlet for much of the period of deposition of the formation. The lake must have extended southward to the present position of Adams Glacier in Adams Valley, and southeast as far as Endicott Gap. Deposits of this lake also occur 4 km up Berg Creek, in Granite Canyon just west of Howling Valley, and on both sides of Seal River 2 km from the mouth. This lake, glacial Lake Adams, was probably formed by damming of the western entrance to Adams Inlet by outwash in Muir Inlet. By 1765 B.P. (Y-304) outwash near Denson Point, 6 km north of Adams Inlet in Muir Inlet, inundated trees on outwash on the west side of Muir Inlet at an elevation of 40 m. At the same location trees on bedrock at an elevation of 56 m were covered by gravel by 1710 B.P. (Y-306).

The thickening of Upper Van Horn gravel in this part of Muir Inlet (Haselton, 1966) may have accompanied the outwash damming of the entrance to Adams Inlet. These gray northerly-derived Upper Van Horn gravels also cover brown gravels of a westerly provenance on the west side of Muir Inlet opposite Adams Inlet. Muir Glacier, which had topped White Thunder Ridge on the west side of upper Muir Inlet by 2120 B.P. (I-1610) was still north of Hunter Cove. It did not destroy vegetation at 61 m elevation at Hunter Cove until 850 B.P. (Y-305).

Meanwhile, outwash from Casement Glacier had been advancing toward Muir Inlet and probably Adams Inlet, although the height of the bedrock between the two is unknown. A progressive increase in elevation of gravels with time is noted away from the front of Casement Glacier in the area of Forest Creek: 2620 B.P. (I-1305), at 28 m, 2175 B.P. (I-88) at 60 m, 1885 B.P. (Y-132-85) at 70 m, and 1800 B.P. (I-1614) at 75 m. These gravels preceded the ice that moved into glacial Lake Adams to deposit the tills.

By 1700 B.P. (I-2687) a lake had formed in Adams Inlet. Glacial Lake Adams was at least 70 m deep shortly after it formed. This is indicated by a date of 1770 years B.P. (I-3069s) on a forest bed at an elevation of 90 m in Granite Canyon. Early in the history of the lake the outlet may have been via Muir Inlet. By 1535 B.P. (I-3151) the lake had deepened sufficiently to deposit material on top of peat at an elevation of 120 m on the southeast side of Casement Glacier. The lake eventually deposited rhythmites at an elevation of 160 m at Endicott Gap (Section 35).

Drainage from the lake by this time may have been by Endicott River. Ice from upper Glacier Bay by now extended beyond Geikie Inlet on the west side of Glacier Bay just south of Muir Inlet. This is suggested by a date of 1540 ± 130 B.P. (Y-4) on wood under till at 91 m on the south side of Geikie Inlet at Fossil Tree Creek. Absence of a nearby glacier precludes deposition of the till by a local valley glacier. The possibility exists that Geikie's Glacier had advanced to this point. A 20-km-wide ice front now reached across Glacier Bay to Muir Inlet, damming a lake in Adams Inlet and the entrance to Muir Inlet (lacustrine deposits occur in the upper part of the section at U.S.C.G.S. Station "Bull" (Goldthwait, personal communication)). Thus, outflow from glacial Lake Adams through Muir Inlet may not have been possible or may have occurred only occasionally by breaking of the ice dam. Deposition of shallow water sands and deltaic gravels with the advance of outwash from Adams Inlet Glacier into White, Adams, and Goddess Valleys eventually destroyed the lake.

The means of containment of glacial Lake Adams at the southern end of Goddess valley is not known. A bedrock high or a deposit of non-glacial material near Endicott Gap may have restricted it to the northern part of the valley.

Relationship to Deposits in Muir Inlet

Deposits of laminated silt and clay that at most places rest on Lower Van Horn (Hypsithermal) gravels were defined by Haselton (1966) as the middle member of the Van Horn Formation. These silty loam lacustrine sediments are rhythmically bedded with individual laminae as much as 5 cm thick, and in many places they contain ice-rafted pebbles. This unit reaches a maximum thickness of 12 m and in most places occurs at elevations of less than 50 m. An isolated deposit of lacustrine material at Canyon Creek in upper Muir Inlet is at 100 m elevation. Radiocarbon dates on wood below lacustrine sediments throughout Muir Inlet range from 4500 - 4200 years B.P.; above the sediments the dates are 2600 - 2200 years B.P. (Haselton, 1967, p. 87).

It is readily apparent from the radiocarbon dates on top of the lacustrine materials that the lake phase in Muir Inlet does not correspond to that in Adams Inlet where deposition of the major lacustrine deposits did not begin until 1700 B.P. and may have continued for 600 years. The thickness of lacustrine deposits in Adams Inlet is generally greater than in Muir Inlet; at many places they reach 30 m and the maximum thickness is 66 m. The maximum elevation for the Adams Formation is also higher than for the middle member of the Van Horn Formation in Muir Inlet.

Although the stratigraphic and radiometric data show that the two lacustrine formations are not correlative, what physical barrier or barriers separated them is not immediately apparent. In resolving this question it is necessary to review the problem of multiple lakes in Muir Inlet.

Goldthwait (1963) offered two hypotheses for the origin of the lacustrine deposits in Muir Inlet. He suggested that these deposits were formed either by an ice dam at the entrance to Muir Inlet, thus creating one large lake in Muir Inlet, or by local fans that created numerous lakes on and adjacent to the main outwash plain in Muir Inlet. Later, on the basis of 10 basal radiocarbon dates ranging from 4665 B.P. (I-80) to 3290 (Y-303) and the common occurrence of the deposits in the Inlet, Goldthwait (1966) postulated that there was only one lake in Muir Inlet. An ice dam created by the advance of ice down Glacier Bay contained this lake in Muir Inlet. Haselton's (1966 and 1967) investigations of the stratigraphy in Muir Inlet led him to believe that, "From the stratigraphic evidence in upper Muir Inlet, it is difficult to demonstrate conclusively that a single lake once filled all Muir Inlet and its tributary arms." In upper Muir Inlet at Canyon Creek the lacustrine sediments were observed by Haselton (1967, p. 83) to pinch out in both the east-west and north-south directions. This and other evidence suggested the possibility of two lakes or two episodes of lakes.

Evaluation of the available stratigraphic and radiometric evidence and a new date on a sample of wood from a section investigated by Goldthwait, has led the writer to conclude that at least two different lakes, and possibly more, existed in Muir Inlet between 4775 B.P. (I-80) and 1765 B.P. (Y-303). In upper Muir Inlet (Goose Cove to Canyon Creek) the dates of the lowest lake deposits are consistently between 4700 and 4300 B.P. (4775 B.P., I-80; 4640 B.P., I-1613; 4560 B.P., I-1616; 4330 B.P., Y-302). Elevations for these deposits range from 15 to 50 m. These dates, although ranging over 400 years, could signify one lake in this part of Muir Inlet; however, Haselton's conclusions, based on stratigraphic evidence, suggest that even in the upper part of Muir Inlet at this time there may have been two lakes. A date (4750 B.P., I-124) on wood below lake deposits in Wachusett Inlet is also in the same range of ages; the deposits here and some of those in Muir Inlet may belong to the same lake.

In the central part of Muir Inlet near Hunter Cove (U.S.C.G.S. Station "Denson") a stump in growth position on 1 m of lacustrine clay and covered by 3 m of gravel and 18 m of lacustrine clay is dated at 3290 B.P. (Y-303). This date is taken to mean a separate lake in that part of Muir Inlet south of Wachusett Inlet. Because there are no lacustrine deposits on the opposite side of the Inlet at this point it is not known whether the lake extended across Muir Inlet. Lacustrine deposits to the south of Klotz Hills may belong to this lake stage. At U.S.C.G.S. Station "Quill" the writer has had dated allochthonous wood collected by Goldthwait from beneath 3.8 m of lacustrine material. This material is overlain by brown gravel with a lithology (slate and schist) indicating derivation of the pebbles from the west side of Muir Inlet. The wood gave an age of 2390 years B.P. (I-3398), which suggests a later date than to the north in Muir Inlet for the formation of a lake here.

In support of these arguments, the time span of over 3000 years (4775 B.P., I-80; to 1765 B.P., Y-303) seems too long to accumulate deposits that, where visible, are no thicker than 18 m. In Adams Inlet more than 60 m of lacustrine material was deposited in less than 1000 years. There is no reason to believe that the available supply of lacustrine material in Muir Inlet was any less than in Adams Inlet. It is doubtful if the ice dam needed by about 4700 B.P. in front of Muir Inlet in the single lake hypothesis could have been provided by a glacier in upper Glacier Bay. Even though ice may have been as far south in Glacier Bay as Reid Inlet at this time (4680 B.P., Y-9) it had not reached far enough or become large enough to stretch across the widest part of Glacier Bay from Geikie to Muir Inlet (1540 B.P., Y-4).

These lakes in Muir Inlet were probably created by outwash fans. The possibility that some of the lacustrine deposits in lower Muir Inlet were formed in lakes dammed by outwash gravels from the west side of Muir Inlet (Morse River) has been suggested by A. T. Ovenshine (Haselton, 1967, p. 87). Continued deposition of outwash in Adams Inlet may eventually create a lake there; this may be analogous to what happened in Muir Inlet several thousand years ago.

The multiplicity of the lakes in Muir Inlet, the outwash-dammed lakes rather than an ice-dammed lake, and the low (less than 50 m) elevations of nearly all the lacustrine deposits explain why similar contemporaneous deposits were not formed in Adams Inlet. The divide between Adams Inlet and Muir Inlet from the Klotz Hills to Casement Glacier is currently between 55 and 60 m, and Van Horn gravels may have been higher, at least during the later lake episodes in Muir Inlet. The Klotz Hills-Mount Wright area may have been filled with outwash gravel; however, a river draining Adams Inlet, in which alluvial gravels from tributary valleys were being deposited at low elevations (ca 10 m at Section 58) existed from at least 3710 B.P. (I-164) until 1980 B.P. (I-2394). At this time a lake of short duration was formed; it may have been very limited in extent and was the forerunner of the glacial Lake Adams.

Berg Formation

Definition

Gravels and sands of unit 5 of the standard section are here named the Berg Formation, and Section 29 on the north side of Berg Creek is designated the type section. The top of this section, at an elevation of 265 m, is 6.2 km N 35° E of the top of Tree Mountain; coordinates for this section are 58° 54' 39" N, 135° 45' 24" W. This formation represents a complex of outwash, deltaic, and fluvial plain sediments. In the type section and in most of the other exposures, the formation is comprised of two parts. The lower member, which includes almost half of the formation at the type section, is mainly fine- to medium-grained stratified sand with some gravel and coarse sand units. The upper member consists primarily of coarse stratified gravel. The Berg Formation overlies the Adams Formation; the base of the Berg Formation is taken as the uppermost occurrence of rhythmically bedded fine-grained sands and silts, clays, or till. Because the lower part of the Berg Formation in many places consists of stratified fine-grained sand and silty sand, the contact is not everywhere sharp. In most places the Berg Formation is overlain by the Glacier Bay Formation, but at Section 69 lacustrine clays with several till units rest on coarse gravel of the Berg Formation.

More than half of the Berg Formation consists of medium- to coarse-grained sand that filled the tributary valleys to the south and east of Adams Inlet. The lesser part of the Berg Formation consists of sands and gravels somewhat similar to the Van Horn Formation in Muir Inlet but differs in that it is composed mainly of steeply dipping foreset beds capped by nearly horizontal topset beds.

Distribution and Thickness

The formation is exposed on all sides of Adams Inlet, but is best developed in the valleys adjoining the Inlet. In Berg Valley it is 130 m thick, in nearby Section 58 at the mouth of Goddess river it is 115 m thick, and in Adams and White Valleys it reaches thicknesses of 100 m and 110 m, respectively. Thinner deposits are present on the north side of Tree Mountain, the entrance to Granite Canyon, and the north side of the Inlet. Near the entrance to Adams Inlet (Sections 12 and 19) the formation consists of 15 to 20 m of medium to coarse gravel and overlies lacustrine silts and clays or fine sands. Gravel deposits as thick as 5 m on the west side of Seal River, underlain by rhythmically bedded silts and clays, and overlain by Glacier Bay till might belong to the Berg Formation. The Berg Formation has been traced as far south as Section 38 (Fig. 2) at Endicott Gap, where the upper and lower members have thicknesses of 53 m and 16 m respectively. Here the base of the Formation is 155 m above sea level. North toward Adams Inlet the base is at an elevation of about 90 - 100 m and in the western part of the Inlet it is at approximately 60 m.

Nature of Deposits

Lower Member

The lower member of the Berg Formation is mainly medium- to coarse-grained sand with some channel gravel and fine sand and silt units (Fig. 20). Gravel layers and bands of pebbles occur with cross-bedded fine- to medium-grained sand. Concentrations of small fragments of wood and occasional seed cones and clay balls occur along bedding planes in

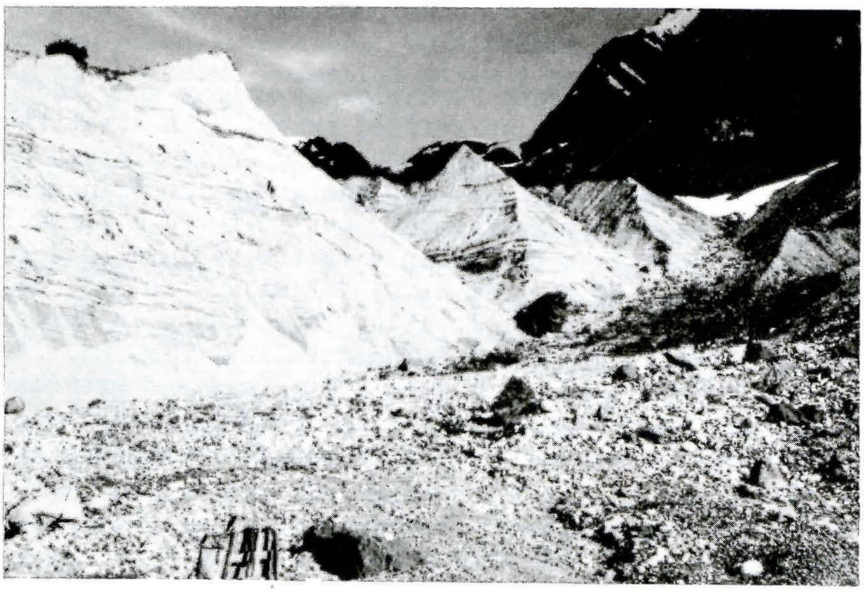


Figure 20. Berg Formation on the south side of Adams Inlet at Section 21. This section is mostly fine- to medium-grained cross-bedded sand with some channel gravels, and is topped by 10 m of coarse cobbly gravel beneath Glacier Bay till.

this unit. The maximum thickness of this member is 70 m and occurs in Berg Valley. The lower part of this member at Section 58 contains lenses of fine sandy gravel in coarse sand and fine gravel. These pockets of gravel are exposed on two faces at 90° to each other and are approximately 1.5 m × 6 m. They occur in coarse sand 5 m thick, that overlies 5 m of generally undisturbed laminated silt and clay, and underlies medium gravel more than 2 m thick. Beneath the lenses of gravel in this feature there is about 1 m of undisturbed sand; on top of the lenses there is as much as 3 m of contorted sand with pebbles. It is suggested that this feature is the result of collapsing ice blocks--remnants of Adams Inlet Glacier that deposited the till units in the underlying Adams Formation.

Upper Member

The upper member of the Berg Formation is mainly gray and grayish-brown medium- to coarse-grained stratified gravel with stratified coarse-grained sand units as thick as 60 cm. The sand units contain pebbles and layers of gravel. Both the sand and the gravel contain driftwood. Much of the bedding is nearly horizontal, but 10 to 20 m below topset bedding are foreset beds that dip steeply up the valleys adjacent to Adams Inlet. At Section 50 the dip is S 3° E at 22° to 26°, and at Section 51 the dip is S 68° E at 12° to 15°. This change in depositional direction reflects the infilling of Pyramid Valley (Fig. 2). At Section 68 in Adams Valley the dip is S 15° E at 27°. Logs and branches occur sporadically throughout the upper member of the Berg Formation, and gravel-filled channels are also visible in the finer-grained gravel.

Pebble Lithologies

Pebble counts were made on 14 samples from the two members of the Berg Formation in Adams Inlet. Plutonic igneous (29 percent), dike rocks (25 percent), and graywacke (21 percent) were the most abundant rock types. Averages of fewer samples from Adams Valley and Granite Canyon (Table 2) were similar. Five samples from Goddess valley showed an average of only half as many plutonic igneous rocks as the averages for the valleys. This difference reflects derivation of material from local bedrock.

The average PI/M ratios for the Berg Formation are very similar (0.74 to 0.87) for all areas except Goddess valley, where the ratio is only 0.35. In Adams Inlet the higher PI/M ratio of the Berg Formation can be used to differentiate it and the Lower Van Horn gravels (0.11).

The Berg Formation in Adams Inlet has an average PI/M ratio of 0.74 which is close to that of the underlying Adams till (0.69) in Adams Inlet. This similarity is expected because the glacier that deposited the Adams till also supplied the outwash material of the Berg Formation. In Goddess valley the source areas for pebbles in the Berg Formation and the Adams and Glacier Bay tills are probably the same, because the average PI/M ratios for these formations are essentially the same (Table 2).

Age and Origin

The only date from wood within the Berg Formation is from an allochthonous sample from topset gravels in Section 69 in Adams Valley, approximately 5 m below Glacier Bay till. This date, 1750 ± 100 years B.P. (I-3150), is about the same as the 1700 B.P. date on autochthonous wood from the base of the underlying Adams Formation at Section 58. The tree from which sample I-3150 came was probably killed by advancing Neoglacial ice, or by formation of Lake Adams, and subsequently reworked to the higher stratigraphic position. The Berg Formation is undoubtedly younger than either of these dates. Rhythmic clays, if considered as varves, of the underlying Adams Formation indicate a minimum of 600 years for deposition of the Adams Formation. Thus adding this time to the date for the beginning of the Adams Formation gives a maximum age for the base of the Berg Formation at about 1100 B.P.

The upper limit to the Berg Formation may be estimated by utilizing rates of outwash buildup determined by Goldthwait (1963) in nearby Muir Inlet and Wachusett Inlet. There Goldthwait inferred from radiocarbon dates a maximum rate of outwash accumulation of 5 m per century (average 1.4 m per century), with an estimated maximum rate of 10 m per century for material deposited near wasting glaciers. Even this estimated maximum rate appears slightly low, for it provides 1000 years for deposition of the 100 m of outwash sand and gravel of the Berg Formation at Sections 58, 50, and 71. This would bring us to within 100 years of the present, with no allowance for the advance of ice that deposited the Glacier Bay till. If we assume that the maximum advance in the Endicott Gap region was at the same time as the Bartlett Cove maximum (250 years ago, Goldthwait, 1963, p. 45), this gives a time for deposition of this material that is too large by 15 percent. This discrepancy may be due to variations in the rates of deposition of the sand and gravel members of the Berg Formation, and the silts and clays of the Adams Formation. Different rates of deposition may exist for sands and dipping gravels of the Berg Formation and the nearly horizontal Upper Van Horn gravels used in Goldthwait's (1963) determination. It may also reflect a slightly later advance of ice to the terminal position at Endicott Gap.

The Berg Formation originated as outwash filling Adams Lake from the north. Fine-grained sand deposits of the lower member covered the lacustrine clays and silts of the Adams Formation. Some ice blocks, stranded in the shallows of the lake and on outwash plains, were covered by sands and gravels. Subsequent melting of these blocks produced disturbed bedding as at Section 58.

During deposition of the sand and the overlying gravel of this formation the ice margin probably stood north of Adams Inlet--possibly along a line from Granite Canyon to the entrance of Adams Inlet (Fig. 2).

The edge of the ice during much of Berg time is not known. It may have been as far south as the north side of Tree Mountain because the

Berg Formation is not present in the central part of the Inlet. The formation may have been deposited in the middle part of the Inlet and later removed by ice as Goldthwait (1963) suggested; however, the available time of about 300 years does not seem long enough. Also the thick Berg Formation in Adams Valley was overridden by Adams Glacier with little apparent erosion.

Gravel filling of lake-filled valleys adjacent to Adams Inlet followed deposition of the finer sediments, as is shown by the deltaic foreset beds in Adams, Goddess, and Berg valleys. Infilling of these valleys probably continued during the final advance of Neoglacial ice over these outwash plains. In White, Adams, and possibly Berg Valleys the advancing outwash deposits and eventually the ice were met by glaciers moving down the valleys.

Glacier Bay Formation and the Last Glaciation

Definition

Glacier Bay Formation is the name proposed by Haselton (1966) for the youngest till in Glacier Bay National Monument. In the type area of Muir Inlet it is as much as 30 m thick and occurs at elevations as high as 610 m. In most places it rests on and contains material incorporated from the Van Horn Formation. The till is a gray bouldery to pebbly loam showing little oxidation and no leaching.

Restriction of the term Glacier Bay Formation to till formed during the last glaciation and forming today leaves out several types of widespread contemporaneous sorted glacial deposits from the stratigraphic classification. For this reason, and because these sorted deposits, which are formed in contact with glaciers, are closely related in time and space to the till, it is here proposed that the term Glacier Bay be extended to include both sorted ice-contact and unsorted glacial drift. Thus unsorted, glaciofluvial, and glaciolacustrine deposits will all be considered as part of the Glacier Bay Formation. This formation is unit 6 of the composite section and is discussed according to the three types of drift. The expanded definition of Glacier Bay Formation would include the "ablation moraine" and some of the "upper gravels" described by Price (1964) in front of the Casement Glacier.

Unsorted Deposits

Distribution and Nature of Deposits

Unsorted drift occurs mainly as a mantle over the other unconsolidated deposits in the lowlands and valleys and varies in thickness from a few decimeters to 10 m. Till was observed in sections at elevations of 165 m, and ground moraines of Glacier Bay age have been interpreted from aerial photographs at elevations of as much as 900 m. The ground moraine is believed to have come from either basal, englacial, or supraglacial drift. Although the distinction may not always be valid, dense, gray, loamy till is considered to have been derived from basal load, and the boulders and cobbles, in many places of different lithology, on top of till and sorted drift are considered to have been derived from englacial and supraglacial loads. Samples used to determine the physical, lithologic, and mineralogic characteristics of the Glacier Bay till were taken only from loamy material.

In addition to the ground moraine there are also a few lateral, medial (Fig. 21), and end moraines. Lateral moraines occur on the northwest side of Granite Canyon at approximately 750 m, and on the north side of Mount Case at 230 m. Near the entrances to Granite Canyon and Adams Valley are bouldery and cobbly moraines as much as 3 m high. A small end moraine is present at 720 m (surface) at the mouth of the first cirque north of Endicott Lake (Plate I). The terminus of Girdled Glacier is marked by a small ridge that may be landslide material in part.

Physical and Lithologic Characteristics of the Till

In Adams Inlet the Glacier Bay till is gray, pebbly and sandy. The average of fifteen mechanical analyses (Table 1) is 55.3 percent sand, 35.9 percent silt, and 8.8 percent clay. This average grain size distribution is a sandy loam. Two samples (71-1, 69-8) of Glacier Bay till fall within the area of the Adams till on the grain-size distribution diagram (Fig. 8). Sample 69-8 is from till overlying lacustrine material of the Glacier Bay Formation and because of incorporations of these clays into the glacier, the till has a mechanical composition similar to that of the Adams till. The similarity of sample 71-1 to Adams till suggests that the 11-m unit from which this sample was taken may also be reworked lacustrine material. A lake may have existed in the hollow of the mountain opposite Camp Adams (Fig. 2).

To determine the significance of the differences in mechanical analyses of the formations, Student's _t_ test was used (Griffiths, 1967, p. 328). If the probability was 0.05 or less, the differences in the units were considered to be real; if the probability was between 0.05 and 0.20 the differences may be real; and if the probability was greater than 0.20 the differences in the two formations being tested are insignificant (Folk, 1961).

In the sand fraction of the mechanical analyses, differences in the means of the Glacier Bay and Adams Formations are significant ($P < .001$); however, the differences in the means of this fraction for the Glacier Bay and Granite Canyon tills are probably not significant ($P \simeq .02$). Comparison of the clay fraction of those latter two formations indicate

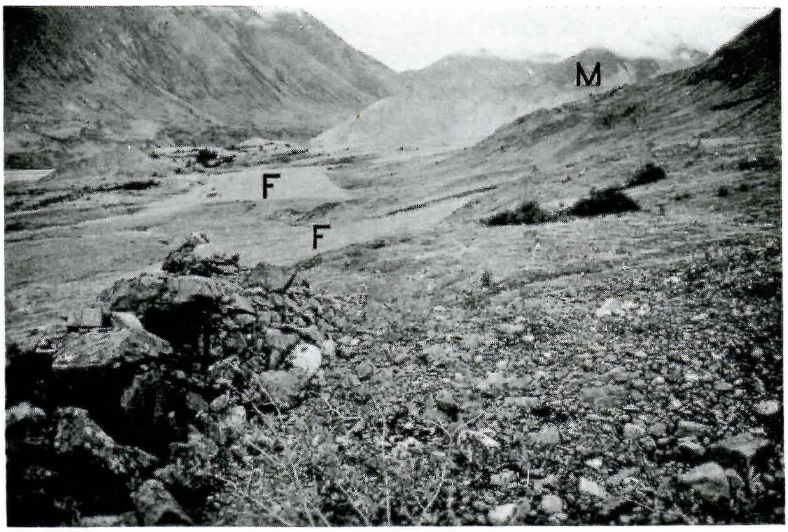


Figure 21. Medial moraine of sedimentary rocks derived from west side of Adams Valley. Note alluvial fans (F), and "Adams mesa" (M) which consists mainly of Berg sand and gravel.

a significant difference, with the probability that the sample came from the same population being less than 1 in 1000. The Granite Canyon till is also distinct from the Adams till ($P < .001$).

Differentiation of the tills on the basis of mechanical analyses is possible and areas of occurrence may be outlined on a triangular coordinate diagram (Fig. 8). These delimited areas give the approximate mechanical composition and have no statistical significance.

The average PI/M ratio of 13 pebble samples from within the nearby area of Adams Inlet is 0.44. This is about midway between the ratios of the Granite Canyon (0.02) and Adams (0.69) tills. Large differences in the PI/M ratio exist between the Glacier Bay till in Adams Inlet and that in the tributary valleys. When all samples of Glacier Bay till are compared to all samples of other tills in the area, the differences are not always significant. With Student's $\underline{t}$ test the Glacier Bay and Adams tills are not significantly different ($P \simeq 0.46$); the Glacier Bay and Granite Canyon tills are different ($P = 0.005$). Thus, this ratio would not be suitable for differentiation of the two most abundant tills--Glacier Bay and Adams. The close stratigraphic relationship of these two tills, separated by the Berg gravels, is also exemplified in their relationships of PI/M ratios with the Berg Formation. The Glacier Bay-Berg comparison shows no significant difference ($P = 0.38$) in these ratios, nor does the Berg-Adams comparison ($P = 0.85$). By PI/M ratios the Berg Formation is more closely related to the Adams Formation. Deposition of the Glacier Bay till from the same glacier that, in a retreatal phase deposited the Berg Formation, and in an advancing phase deposited the Adams till, accounts for these similarities in pebble composition.

Comparison of the mechanical and lithologic analyses of all the samples of Glacier Bay till in Adams Inlet with the same till in Muir Inlet indicates that they are closely related. The average mechanical compositions are almost identical (Table 1). Student's $\underline{t}$ test on the mean sand percentages gives a probability of 0.78 that the samples are from the same population. The average PI/M ratios for the two localities are very close. For Muir Inlet the ratio is 0.67; for Adams Inlet the ratio is 0.70. Comparison of the two means using Student's $\underline{t}$ test shows a high correlation ($P \simeq 0.9$).

Clay Minerals

Two samples of Glacier Bay till, one from Granite Canyon (66-2) and one from the east end of Adams Inlet (74-1) 2.5 km to the southeast, were analyzed for clay minerals (Table 3). Illite and chlorite are dominant with illite being about twice as abundant as chlorite. Little difference exists in the clay composition of those two samples, suggesting uniformity in the till. For purposes of differentiation of tills in the area clay minerals may be suitable only in the case of the Adams Till (Table 3). The Glacier Bay and Granite Canyon tills have about the same clay mineral composition.

Elemental and Carbonate Analyses

The percentages of elements in both samples of Glacier Bay till are approximately the same, again suggesting uniformity over short distances in this till sheet (Table 4). Much closer agreement exists between these two samples than between the two samples of the Adams Formation (Table 4); however, the mechanical composition is not the same for the two different units of the Adams Formation, which may account for the compositional differences within the formation.

The average values of Ti, Fe, Mn, and K for the Glacier Bay till are significantly different from values for the Adams till. In the HCl-treated samples the Zr values are also different. Comparison of the Granite Canyon till (unweathered) with the Glacier Bay till shows that there are also significant differences in percentages of Ti, Zr, Mn, and K, particularly in the HCl-treated samples.

With HCl leaching, both Glacier Bay till samples showed compositional changes similar in magnitude and direction to other units. One difference was found that distinguishes this till from other tills. The percentage of Zr increases in the samples subjected to leaching. This is the reverse of the changes in the unweathered Granite Canyon till and Adams till. It may thus be possible to use absolute changes or ratios of several elements upon leaching, in characterizing a till sheet. Samples are too few in this study to warrant conclusions on such calculations.

Carbonate analyses of the -200 mesh fraction average 7.8 percent total carbonate (Table 5). This is the lowest average of the three tills, but not very different from the Adams till (9.9 percent). Although only three samples of each till were tested, the calcite/dolomite ratios may serve to differentiate these two tills. For Glacier Bay till this ratio is 2.5; for Adams till the ratio is 6.6.

In the case of the Glacier Bay till it appears from the limited data that several methods may be used to differentiate and characterize tills in this area. Mechanical analyses, pebble lithologies and their PI/M ratios, clay minerals, elemental analyses of the sand and silt fraction, and carbonate analyses of the -200 mesh fraction indicate, with varying degrees of certainty, that separation of some tills is possible. Correlation of the Glacier Bay till in the area of Adams Inlet with that in Muir Inlet is possible using mechanical analyses and pebble lithologies.

Age and Origin

The Glacier Bay Formation is a diachronous unit that has been forming in Glacier Bay for about 4700 years as indicated by the post-Hypsithermal advance of ice in western Glacier Bay (Goldthwait and others, 1966). In Wachusett Inlet deposition of the Glacier Bay Formation began about 2735 years ago (I-122). To the southeast in Adams Inlet the deposition of this unit was even later. Following deposition of the Adams and Berg Formations, ice that deposited Glacier Bay till in this area advanced over these formations. In some areas of Adams Inlet, because of the time needed for deposition of these underlying units, this till may have begun forming as late as 300 years ago as in Goddess valley. This is shortly before the maximum extent of ice in lower Glacier Bay 250 years ago (Goldthwait and others, 1966).

Glaciofluvial Features

Depositional Features

The glaciofluvial deposits included in the Glacier Bay Formation are those ice-contact deposits sorted by running water. This includes eskers, kames, kame terraces and deltas, alluvial fans, and pitted and collapsed outwash. Most outwash is included in the Seal River Formation described later.

Most of these deposits are of gray medium to coarse gravel. Stratification and pebble roundness are variable depending on the feature and the source of the pebbles. Pebbles derived from Glacier Bay till are mostly subrounded because they were in turn derived from the Berg Formation. Pebble lithologies are probably similar to those of the till in the Glacier Bay Formation. Some of these deposits are sandy, as is the kame delta at the eastern end of Adams Inlet near the entrance to Berg Valley, and the top of some eskers in the Casement proglacial area (Price, 1965). Pockets of ablation and flow till are mixed within these deposits, and some scattered angular boulders derived from supraglacial debris rest on the glaciofluvial deposits. In several parts of Adams Inlet buried ice (Plate I) has been observed beneath glaciofluvial gravels of eskers, kames and kame terraces. In other parts of the Inlet the presence of buried ice is inferred from collapse features.

Most of the glaciofluvial deposits are located near the mouths of valleys on the north side of the Inlet near Seal River. Only minor deposits occur on Adams island possibly because of the protecting cover of stagnant ice in the center of the Inlet. Eskers are most numerous along the north shore of Adams Inlet where marginal Casement drainage and mountain runoff proceeded along the side of Adams Inlet Glacier and then under the ice. These are mostly small eskers associated with kames. The longest esker is 2 km and is as much as 4 m high. The largest eskers and those in the esker system at the terminus of the Casement Glacier (Price, 1965). This 1.5-km-long system, believed by Price to have been formed supra- or englacially, contains ridges as high as 37 m, with ice cores 18 m thick.

Kames are most numerous on the north and south shores of the Inlet. On the north shore some of these kames consist of coarse cobbly gravel in mounds 10 m high. Eskers and crevasse fillings are associated with these kames. Similar associations of kames, eskers, and crevasse fillings occur on the east side of the entrance to Adams Valley, at the entrance to White Valley, and between these two valleys. The stagnant or slowly moving ice mass in the Inlet combined with heavily debris-charged outwash streams and mountain streams capable of picking up material from the Berg Formation, provided a situation for formation of ice-contact deposits. The largest kame in Adams Inlet, over 90 m high, is 3.5 km southwest of Camp Adams on the west side of White River (Fig. 12). It is not known whether or not ice remains beneath the gravel. Parts of the esker ridge on the east side of Adams Valley show seepage and may contain buried ice.

A deposit of ice-contact gravel 1.4 km south-southeast of Camp Adams is believed to be a collapsing kame terrace. It is ice-cored, as determined by seismic and visual observations, beneath a cover of about 3 to 6 m of medium gravel. The deposit is covered by *Dryas* and some low alder and willow scrub, and the surface contains kettles, some as deep as 4 m. The higher parts of the surface of the terrace are not pitted and suggest either uniform settling or absence of ice beneath that part of the terrace. The gravels were deposited as outwash from the wasting Adams Glacier that occupied a small col south of the 460-m-high bedrock knob, 2 km to the east (Plate I). This outwash was deposited along the side and on the top of Adams Inlet Glacier.

What may be termed a kame delta, that is, a delta in contact on one side with ice or built on top of ice, is located at the mouth of Berg Valley on both sides of the river. This deposit consists mainly of pebbly fine to coarse sand with occasional angular cobbles on the surface. Several kettles over 10 m deep occur on the surface. Some of the deposit grades into kame terraces toward the valley, and terraces are cut into the kame delta and kame terraces. These deposits are believed to have been formed during a late stage of glacial Lake Endicott. They vary in elevation from sea level to over 120 m. Similar ice-contact deltas and terraces occur adjacent to the westward-flowing streams south of Berg Valley, on the south side of Pyramid river, and adjacent to the "amphitheaters" on the east side of Goddess river 1.5 km south of Pyramid river.

A good example of ice-contact alluvial fans occurs on the south side of Adams Inlet less than 0.5 km west of the large kame terrace south of Camp Adams. These fans were formed after most of Adams Inlet Glacier had wasted or was buried. Runoff from the unnamed mountain to the south eroded the Adams and Berg formations and deposited this material in fans over, and in some cases against, stagnant ice. With melting of these ice blocks the alluvial fans became pitted with small kettles. This process is active today. On the south side of Adams Inlet near Camp Adams a kettle lake that was over 3 m deep in 1966 became filled with gravel when the stream on the adjacent alluvial fan was diverted. In 1967 there was no trace of the kettle. Continued melting of any remaining buried ice will no doubt result in another kettle if the stream changes course again.

Pitted outwash deposits are not common in Adams Inlet. The two main occurrences are on each side of the entrance to the Inlet. The outwash terrace of White Valley that reaches the entrance to the Inlet is pitted down to the water. Outwash probably extended another 500 m into the Inlet at one time and covered stagnant ice blocks there. These areas are now part of the tidal flats. Changes in the topography of the next higher terrace have taken place since 1948. A small kettle containing a lake was enlarged by melting of buried ice with the result that the sea shore extended 250 m inland to destroy the lake. A fault in outwash near this kettle is thought to have formed during the 1964 Alaskan earthquake. K. Loken of Juneau photographed this change shortly afterward. It is the writer's opinion that this feature was formed during the earthquake by the settling of outwash into voids left by melted ice. The earthquake triggered large scale settling that in time would have proceeded to form a kettle.

Pitted outwash also occurs on the north side of the entrance to Adams Inlet, opposite White Valley, at the mouth of an abandoned outwash channel. Outwash from the channel was being deposited around stagnant ice blocks by 1929. Most of the pitted outwash there is now in the tidal flat area. Other small kettles occur in the fan delta of the Adams River, along the west side of Goddess river, and in several places in Granite Canyon.

Erosional Features

Ice-marginal channels are channels formed by meltwater in contact with glacial ice. They can be classified according to their position relative to the ice as lateral, frontal, and subglacial. In Adams Inlet these channels, which may be excavated in bedrock or unconsolidated materials, vary in length from about 100 m as on the mountain on the east side of Endicott Gap, to 5 km as on the north side of the entrance to Adams Inlet. They occur at elevations from a few meters above sea level to 760 m (2500 ft) on Tree Mountain and 730 m (2400 ft) on the northwest side of Granite Canyon. Ice-marginal channels can be used to determine former thicknesses and slope of glaciers (Mannerfelt, 1949). In Adams Inlet, because of an unusual and complete photographic record (Field, 1947; and others), it is also possible to determine, in some cases, when the channels were occupied. Both lateral and frontal channels are discussed below by location.

Casement Area

The 2-km-long channels on the southwest side of Snow Dome (Plate I, Fig. 2) at elevations of 300 m (1,000 ft) and 245 m (800 ft) respectively were probably formed before 1929. The lower channel is close to the Poplar Line vegetational stage (Goldthwait and others, 1966), which indicates deglaciation by 1920 to 1925. The upper one is 60 m higher and, using a deglaciation rate of 6 m per year (described in succeeding chapter), was probably active 10 years earlier. These channels probably supplied meltwater and debris for the formation of the eskers on the north shore of Adams Inlet.

By 1941 the most southerly channel, 1 km north of the Inlet on the east side of Seal River, was functioning (Field, 1947)(Plate I). The sequence of development of the drainage channels on this and the west side of Seal River have been described by Welch (1964). By 1948 the channel (30 m deep) 1.1 km to the northwest was active. This channel and the other two to the north were active until the early 1950's when the present outwash channel was formed.

In 1929, meltwater on the west side of Seal River was carried south to the entrance of Adams Inlet and also west to Muir Inlet. With continued retreat of the Casement Glacier the west tributaries of this south channel were abandoned and are now non-concordant with this channel. In the early 1940's this south channel was abandoned in favor of the curved channel to the north that flowed into Seal River. This channel ceased to function by 1948.

South Side of Adams Inlet

Evidence in the ice margin map of Goldthwait (Goldthwait and others, 1966) indicates that the drainage channels on both sides of the entrance to White Valley were probably formed between 1910 and 1930. Some parts of these channels in unconsolidated material may have been cut subglacially. Although no longer carrying meltwater, several of these deep gullies (Plate I) are now forming alluvial fans.

On the west side of Adams Glacier near Adams Inlet, ice marginal channels in bedrock occur at elevations of 60 m (200 ft) to 300 m (1,000 ft) and are as much as 4 m deep. The highest channel may date from 1890 because the ice was at this level then. Small marginal channels on both sides of the entrance to Adams Valley were formed between 1929 and 1941. The large gullies in the north end of Adams mesa were being formed in 1946 (Field, 1947).

Several small (5 to 10 m long and 3 m deep) marginal channels at 825 m (2,700 ft) on the northeast side of Tree Mountain are close to the elevation of the trim line and were probably formed about the maximum of the last glaciation. On the same ridge of the mountain at lower levels are three lateral channels. These are at elevations of 485 m (1,590 ft), 455 m (1,490 ft), and 418 m (1,375 ft) and are respectively 2, 3.5, and 4.5 m deep. The uppermost one was probably occupied in 1890 as this is the approximate position of the ice at that time (Reid, 1892).

Several marginal channels are present on the north side of Mt. Case (Fig. 22). At 230 m (750 ft) is an eastward-sloping marginal channel (Fig. 23) that is brush-free for about 200 m. Here the channel is about 35 m wide; the outer lip is a moraine of sandy till about 5 m above the floor of the channel, which was partly covered at the time of observation.

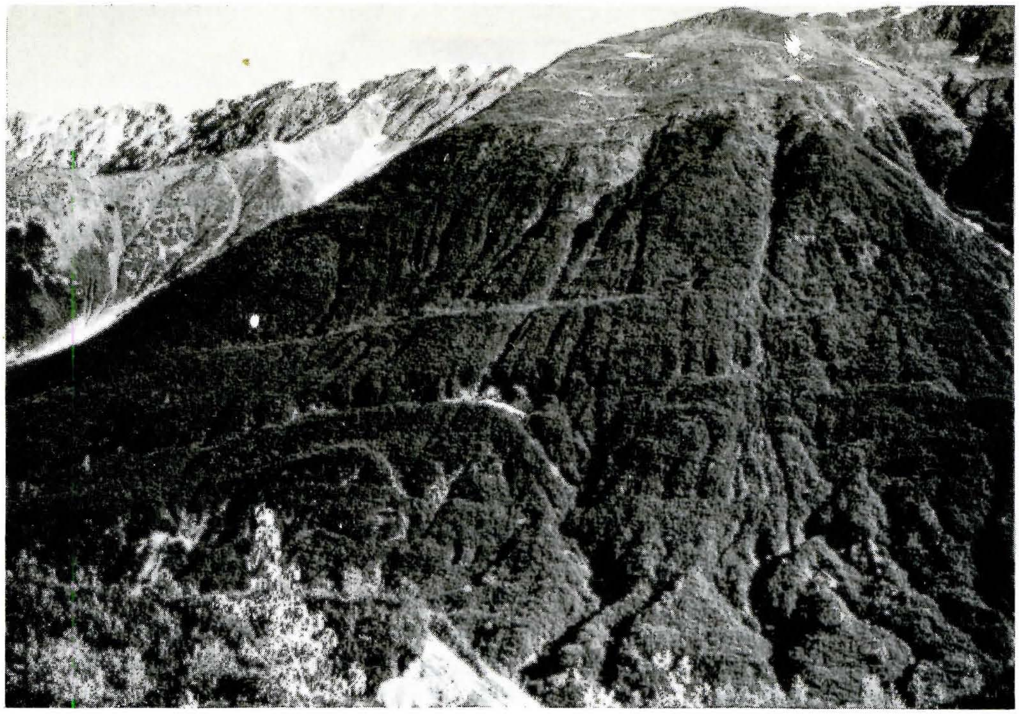


Figure 22. Marginal channels on the south side of the entrance to Adams Inlet (north side of Mt. Case). Lower one is at 230 m (750 ft) and the upper one is at approximately 335 m (1100 ft) elevation.

Figure 23. View eastward toward Tree Mountain from the lower (230 m) drainage channel on the north side of Mt. Case.

The inner, nearly vertical wall of the channel is 30 m high at the highest point. The moraine is fresh and is associated with the most recent glaciation of the area. The channel was probably also formed at the same time; however, the slope to the east does not agree with the observed ice slopes (Reid, 1896) or the direction of ice flow suggested by medial moraines on the surface of Adams Inlet.

The answer may be that these channels developed when the entrance to Adams Inlet was dammed by ice from the Russell drainage (Ovenshine, 1967) in Glacier Bay. Muir Glacier, blocked by this ice at Glacier Bay, began to spill over into Adams Inlet. During a stillstand of this ice, possibly during Berg Formation time, these lateral channels were formed in the north side of Mt. Case. This blocking of Muir Glacier also provided for the building of the extensive Muir Icefield recorded on the early maps.

Granite Canyon and the East End of Adams Inlet

The channels on the northwest side of Granite Canyon between 670 m (2,200 ft) and 730 m (2,400 ft) are just below the lateral moraine marking the margin of the last ice and were formed during the early stages of deglaciation (Plate I). If we assume that the channels were not wholly subglacial and use a conservative ablation rate (also assumed to be constant) of 6.1 m (20 ft) per year (Reid, noted in Field, 1947), these channels, which are 425 m (1,400 ft) above the 1890 ice position, may have been formed as early as 150 years ago.

Other channels in Granite Canyon and near the entrance to Girdled Glacier Valley are much younger. Several frontal channels draining a small marginal lake between these two valleys were photographed in 1941 (Field, 1947, Fig. 12).

The five short bedrock channels, each less than 200 m long, east of the present position of the outlet from Granite River (Plate I) were covered by ice in 1941 (Field, 1947). It is inferred that some of these were formed as subglacial channels draining the stagnant ice mass at the entrance to Granite Canyon. Some of these channels may have been contemporaneous, but there was also a shift to the west with continued erosion of ice and gravel that contained these streams. The present position of Granite River (Plate I) is 400 m west of that in 1948 (Juneau D-6 quadrangle).

The large bedrock channel on the south side of Berg River at 395 m (1,300 ft) was formed subglacially prior to 1890. The smaller and lower channels nearby may have been occupied by subglacial streams draining Berg Lake toward glacial Lake Endicott. Small ponds at the base of two of these channels are probably potholes.

Endicott Gap

Lateral marginal channels are common on both sides of the valley. On the west side they are as much as 700 m long and indicate a southerly flow. Those above 300 m (1,000 ft) were formed prior to 1890; those between 240 m (800 ft) and 300 m (1,000 ft) may have been tributary to glacial Lake Endicott, because they are close to the margin as mapped in 1890.

On the east side of Endicott Gap north of Endicott Lake, a 2-km-long sequence of en echelon lateral marginal channels are cut in bedrock. These channels are generally less than 300 m long and occur at elevations of 300 m (1,000 ft) to 455 m (1,500 ft). In one case, two channels that are separated by more than 0.5 km appear to have once been continuous since they have the same gradient and direction. This suggests that at least the upper channels were formed partially englacially or supraglacially. Many short channels follow the 1,000-ft contour and were probably formed adjacent to stagnant ice prior to glacial Lake Endicott, which was in existence before 1890. The marginal channel between 300 m (1,000 ft) and 395 m (1,300 ft), and 0.5 km south of Endicott Lake, was draining an earlier high stand of a lake at the present site of Endicott Lake. A kame moraine was forming at the entrance to the lake. Outwash from this channel built the high terrace just below 300 m (1,000 ft) on the east side of Endicott River. Frontal ice-marginal channels were formed in outwash when meltwater flowed from the east side of the valley westward through and against the debris-covered ice which later melted to form the kame moraine at Endicott Gap (Plate I). Some of these channels may have drained glacial Lake Endicott.

West Side of Endicott Valley

In the southeastern region of the map area (Plate I) are lateral channels between 425 m (1,400 ft) and 520 m (1,700 ft) in elevation. They formed adjacent to expanded valley glaciers tributary to Endicott Valley probably during the last glaciation. Channels 2 and 5 km east of the southeast corner of the map area were probably formed at the end of the Wisconsin glaciation. A 300-km-wide overflow channel north of Endicott River (off map sheet) is 3.5 km long and was also formed during the latter stages of the Wisconsin glaciation when ice blocked upper Endicott Valley.

Glaciolacustrine Features

Three glacial lakes in the vicinity of Adams Inlet appear on early maps (Reid, 1892, 1896)(Fig. 3). Field (1947) noted two other lakes in Granite Canyon, one of them apparently not ice-dammed (Fig. 24). Several of these lakes, for which there exists an historical record, have left geologic evidence as to their position. In at least two valleys (White, Adams) there is evidence of ice-dammed lakes prior to the Neoglacial

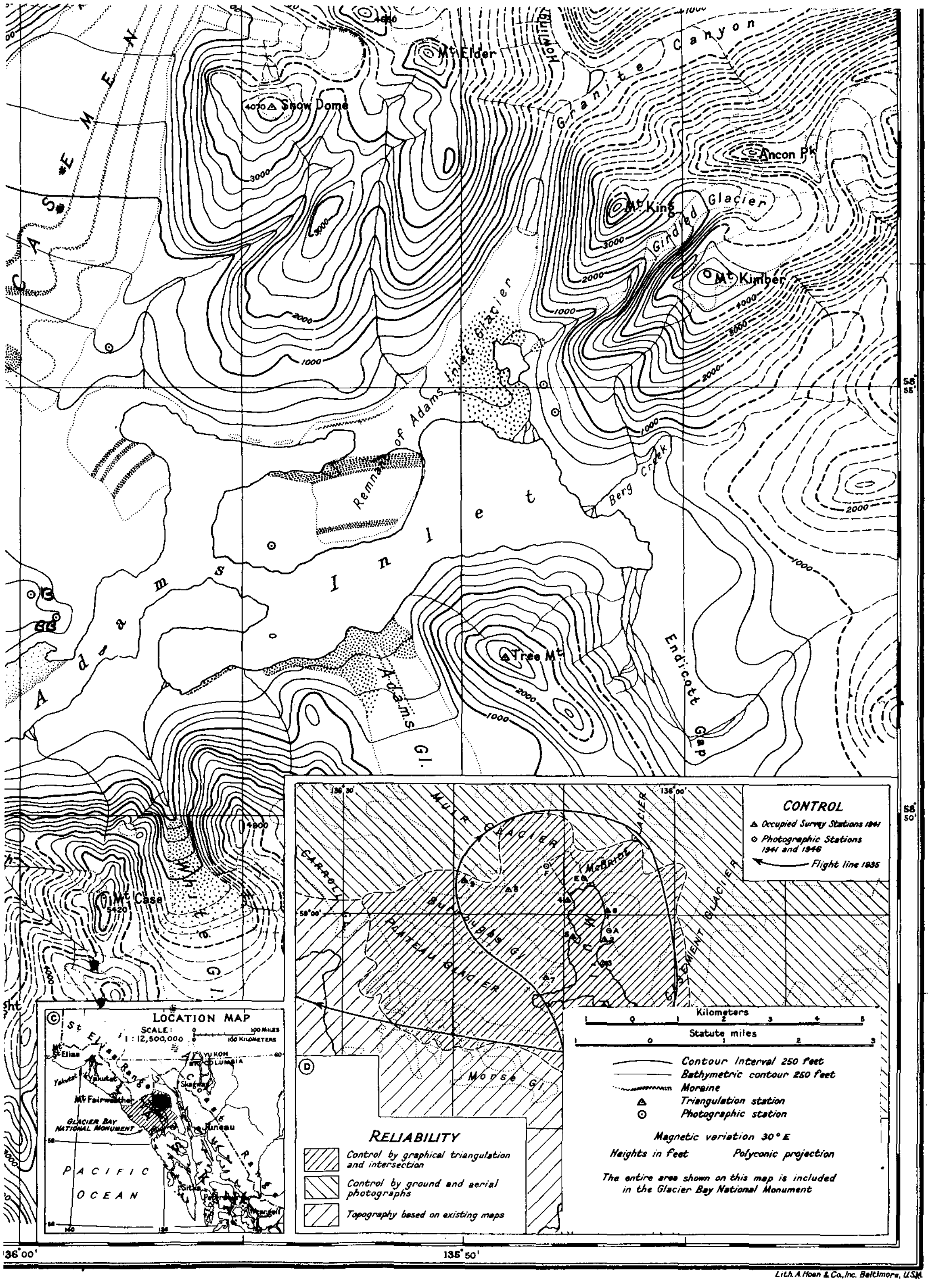


Figure 24. Map of Adams Inlet 1941-46 showing ice margins by W. O. Field (1947). Taken from a larger map of Glacier Bay National Monument; the explanation is from another part of the map. Map courtesy of Geographical Review.

maximum. A brief account of these two pre-glacial lakes is given below, followed by an account of the glaciolacustrine features of the better documented post-Neoglacial Lakes.

Early Neoglacial Lakes

The two lakes for which stratigraphic information exists are in Adams (Section 69) and White Valleys (Section 76). Adams Valley lake was formed by ice from Adams Inlet advancing southward over the Berg Formation. At Section 69 the gravels at about 200 m elevation are overlain by 1.7 m of laminated gray silt. Other silt and fine sand layers occur above the overlying till and may also be of lacustrine origin. Neither the outlet nor the size of this lake are known. The lowest divide on the map (Fig. 2) is 490 m (1600 ft); drainage may have been subglacial.

Evidence of the other ice-dammed lake is the 26 m of laminated gray silt and clay above an elevation of 231 m in Section 76. Rhythmites 5 cm thick and containing pebbles are common. The high elevation (267 m) of deposits suggest that this was a local valley lake. It was dammed by ice in Adams Inlet probably during and after deposition of the Berg Formation. Earlier it may have been joined to the late stage of Adams Lake.

Late Neoglacial Lakes

Granite Canyon

Granite Canyon was the site of at least three different lakes according to maps by Reid (1896) (Fig. 3) and Field (1947) (Fig. 24). In 1892 a small lake occupied Howling Valley possibly to a level of 455 m (1,500 ft). Little else is known about this lake. Aerial photographs give little indication of a shoreline. A self-draining lake of the same size is present on the northwest side of Casement Glacier north of the map area. This lake and its drainage mechanism has been described by Lindsay (1966).

Two other lakes in Granite Canyon appear on Field's map (Fig. 24). Each of these has an area less than 0.6 km^2. The smaller one occurs in the middle of Granite Canyon near the entrance to Howling Valley. It is possible that the lake was much larger at one time, occupying the upper portion of Granite Canyon, and dammed by ice near Adams Inlet. No evidence for this was found however. This lake had disappeared by 1948.

The other lake in Granite Canyon was located about 1 km from Adams Inlet near the valley of Girdled Glacier. More than 60 percent of the shoreline of this lake was debris-covered ice. Water drained through bedrock gorges to Adams Inlet (Field, 1947, Fig. 12). This lake had also disappeared by 1948. A fine to medium pebbly sand, part of a deltaic deposit, is now present in part of this lake area. By 1966 a narrow, 500-m-long lake (Plate I) occupied the site of the former ice-dammed lake, apparently formed due to collapse of buried ice blocks.

Berg Lake

Named for the numerous icebergs that floated in it, Berg Lake once occupied an area of 3.1 km^2 (Fig. 3) in the lower part of Berg Valley. For much of the life of the lake, drainage was southward through, under, or along the ice to glacial Lake Endicott.

A 10-km-long linear feature occurs on the south side of the valley at about 260 m (850 ft) (Plate I). This is believed to be a shoreline of Berg Lake. At the entrance to Berg Valley, drainage channels at 275 m (900 ft) and 365 m (1200 ft) may mark higher stages of Berg Lake. Pothole lakes occur at the ends of these channels (Plate I). Terraces between 245 m (800 ft) and 275 m (900 ft) in Berg Valley may be remnants of valley train deposits that terminated at Berg Lake (Fig. 3).

Glacial Lake Endicott

Terminology:--The term Endicott Lake is now applied to a small lake on the east side of Endicott Valley at the head of Endicott River. According to Orth (1967) this lake was originally called Main Lake by Reid in 1890, after the valley named by John Muir in 1882. This is erroneous however, because Endicott Lake, which is outside of the Monument and not in it as Orth (1967, p. 314) reports, was never part of a nearby larger glacial lake that occupied the valley between Endicott Gap and Adams Inlet (Fig. 3). The term Main Lake was used for this glacial lake by both Reid (1892) and Cushing (1891) and is on their maps. It is separate from the small lake (not shown on Cushing's map) to the south on the east side of the valley. The names of the glacial lake and the valley it occupied were later changed to Endicott by Reid (1896). This term was used by geologists and others until the lake disappeared (about 1930). Between that time and the publication of the 1949 quadrangle map the small lake inherited the name Endicott. This switching of the name to a different and existing lake introduces confusion in the present discussion. It could be avoided by using the term Main Lake for the glacial lake; however, Endicott was the official name for the glacial lake before its disappeance. For this reason, and its earlier popular usage, it is believed that the name should be retained for the glacial lake. Thus, when referring to the glacial lake, the term "glacial Lake Endicott" will be used herein; the small existing lake will be referred to as Endicott Lake.

Stage I:--Glacial Lake Endicott was the largest of the post-Neoglacial lakes in the area. In 1890 it had an area of 10 km^2 at an elevation of about 250 m (820 ft). Several indicators of this former lake level are known. Two deltas, one near the entrance of Pyramid valley (Fig. 25) and the other on the opposite side of Goddess valley (Plate I), are just above the 800-ft contour line. Another indication of the former lake level are the abandoned outwash channels along the National Monument

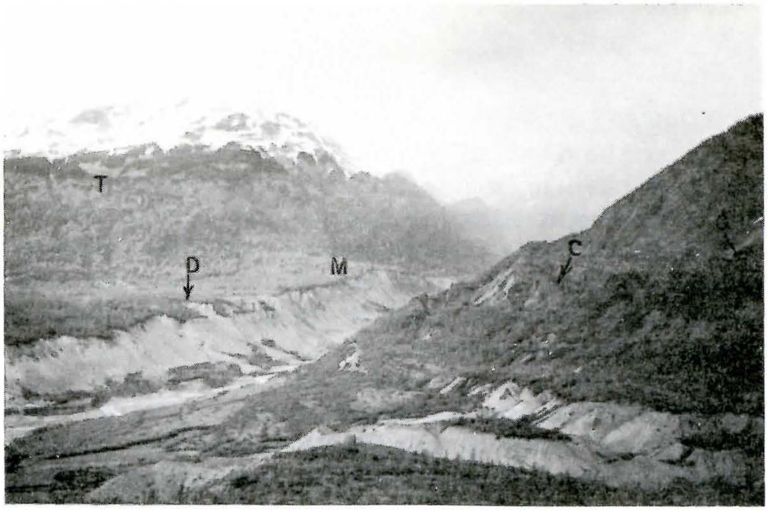


Figure 25. Delta (D) of glacial Lake Endicott at an elevation of 250 m (820 ft). Composite moraine (M) of gravel and till, and lateral channels (C) are both between 305 m (1000 ft) and 335 m (1100 ft). Trim line (T) is at 655 m (2100 ft). Viewed eastward up Pyramid Valley.

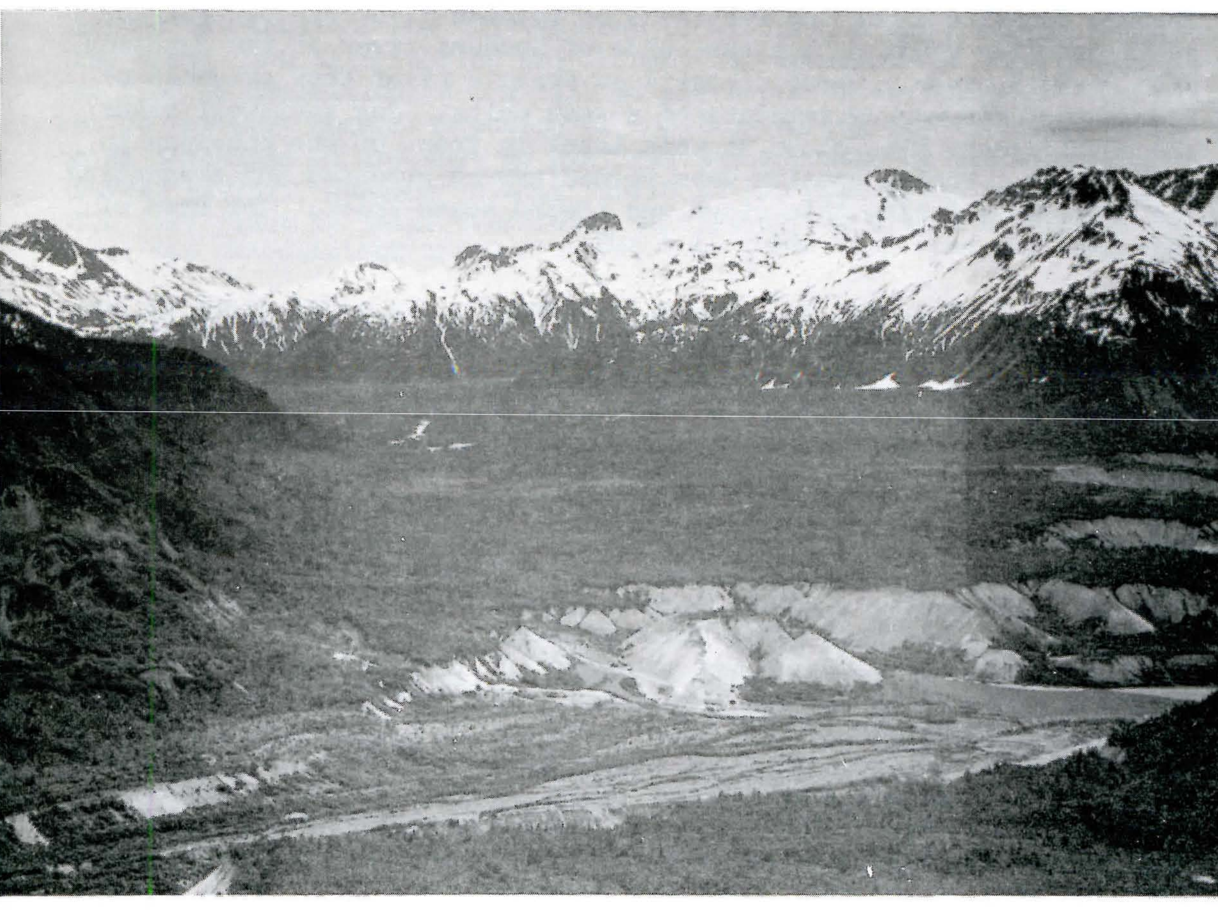


Figure 26. View southward through Endicott Gap with Pyramid Valley entrance at lower left. Glacial Lake Endicott occupied depression on left and extended to the right beyond the edge of the photograph. Berg Formation gravel and sand occurs in bluffs.

Boundary (Endicott Gap, Fig. 26) between Goddess and Endicott Valleys. The lowest point is on the west side and is just over 245 m (800 ft). Evidences for what may have been an earlier higher stage of the lake are drainage channels on the west side of the divide at elevations above 275 m (900 ft)(Plate I). These channels terminate in a kettle depression that was part of the outwash plain in 1892 (Fig. 3). The channels may be remnants of late-stage outwash from the glacier terminus, prior to formation of glacial Lake Endicott.

Glacial Lake Endicott doubled in area compared to the size in 1892, before the termination of Stage I. Calving caused the ice dam to retreat at least to the south side of Tree Mountain where one of the high deltas was built. By this time Berg and Endicott may have become one lake. Lacustrine deposits of this lake consist of fine sands, silts, and clays as much as 5 m thick. Some sandy deposits near the center of the lake may be detritus washed out of, and deposited against, a stranded iceberg. Raised isolated channels as long as 400 m are present in the middle of kame deposits (Plate I). They are remnants of a collapsed outwash plain.

Stage II:--The second stage of the lake is thought to have had an elevation of about 90 m (300 ft) to 120 m (400 ft) as indicated by the terrace deposits along the west side of Goddess river and the sag and swell gravelly sands on the east side of Goddess river at the junction with Pyramid river. These deposits on Pyramid river show evidence of ice contact and may be remnants of a kame delta. Sag and swell sands and pebbly sands also occur at this level in front of the entrance to Berg Valley. These deposits were formed around stranded and stagnating ice. Drainage during this stage was toward Adams Inlet. The area of this lake may have been about 11.6 km^2.

Stage III:--As was the case with Stage II, the third stage is not well defined in elevation or areal extent, possibly because of the rapidly wasting ice dam and resultant variation of lake level. An indication of the elevation of the lake is given by two pitted fan-deltas on the west side of Goddess river near Adams Inlet (Plate I). These are at elevations of 20 - 30 m. At the east end of Adams Inlet at a slightly lower elevation is an area of pebbly sand that may be a collapsed delta. Kame gravels, perhaps derived from bluffs containing the Berg Formation, lie at about 30 m elevation between Berg and Pyramid valleys on the east side of Goddess river. These gravels have a vertical range suggesting that they may have been formed in part during Stage II. Abandoned channels, kettles, and potholes are associated with these deposits formed near the edge of this lowest lake stage. Silt with striated pebbles and cobbles was deposited in this lake. At Section 7 these silts are 70 cm thick; across the river they are clayey and are 25 cm thick. Drainage during this stage was toward Adams Inlet. A 1929 U.S. Navy photograph of Adams Inlet (Field, 1947, Fig. 9) shows the iceberg-filled lake during this stage. By 1935 the last stage of this lake had been drained (Field, 1947, p. 382).

Adams Inlet is situated in such a position that reversals of ice-flow directions in this basin were possible. Two methods of filling the Inlet with ice are visualized: overflow of ice from Muir Inlet and Casement Glacier, and valley glaciers from south and east of the inlet. Outflow from Adams Inlet is possible through Endicott Gap or through Muir Inlet. The resultant flow directions at any one time depend on a combination of factors including differences in precipitation between east and west parts of the Monument, response times to changes in precipitation, rate of wastage of ice (including differences in tidewater and grounded glaciers), elevations of catchment basins, and slope of the glacier bed. Besides the historical account of deglaciation, information on ice movement is available through till fabrics, striae, ice-sculptured bedrock, and trim lines.

Till Fabric

Most of the till fabrics in Glacier Bay till were on material a meter or more beneath the surface of the exposure; however, because of limited thickness of the unit at some localities, a depth of 0.5 m was also used. The till fabrics are thought to represent the flow of ice at the time of deposition of the till at the depth of the fabric. This may not be the most common direction of ice flow during deposition of the total thickness of the unit as has been demonstrated by Dreimanis and Reavely (1953). Changes in till fabric in the underlying material may also take place by later overriding of till by ice from a different direction (MacClintock and Dreimanis, 1964).

Fabric maxima for Glacier Bay till are marked on the directional features map (Fig. 10) with 'GB.' The fabric diagrams for this till are given in Fig. 27 in which the numbers refer to sections or stations at which the fabrics were done. Locations for the sections are given in Fig. 2; locations of the stations are described in the text.

Fabric 66-3 GB (Fig. 27) is from the recently deglaciated area on the northwest side of Casement Glacier and indicates a direction of flow parallel to striations on top of the recently exposed nunatak 300 m to the southeast (Fig. 10). The fabric diagram shows a slight dip in the upglacier direction. Fabric diagram 66-9 GB is from a cut in front of Camp Adams, and shows a definite maximum at S 80° E. A large striated and elongated boulder on the beach has a long axis at S 60° E; striae and grooves nearby have a bearing of N 83° E. The fabric suggests flow out of Adams Valley as does the boulder orientation, and the striae represent a flow direction that more nearly parallels the medial moraine on the island.

The three fabrics measured on the west end of Adams island are not as simple as the other fabrics in Adams Inlet. Fabric 67-42 GB was

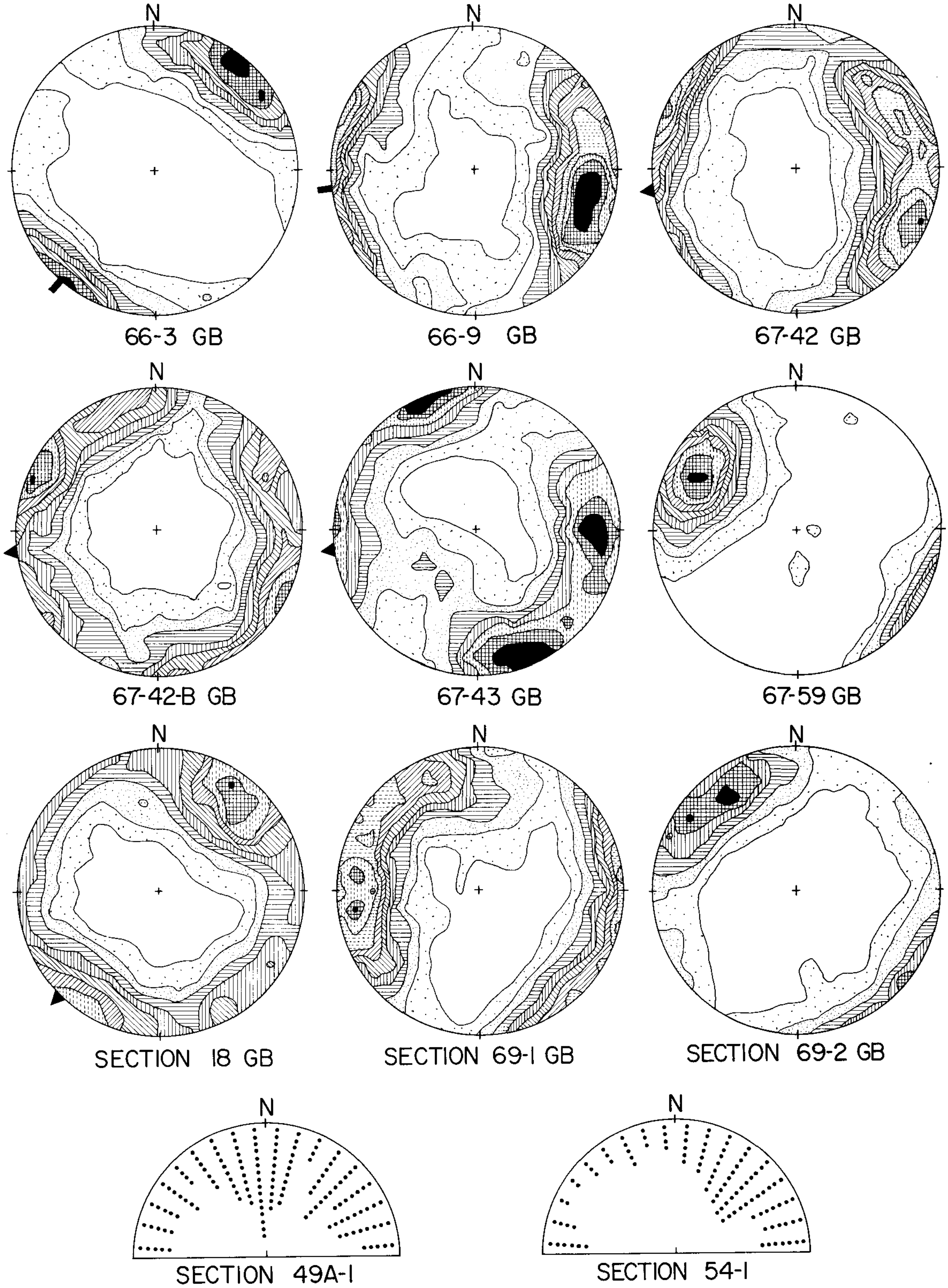


Figure 27. Till fabric diagrams of Glacier Bay till (GB and Section 54-1) and one sample of Adams till (Section 49A-1). White area is the lowest concentration. Contour interval is one sigma except for 66-3 GB, 67-59 GB, and Section 69-2 GB which are two sigma. Bar indicates ice flow direction as determined from nearby striae; triangle indicates flow as determined from surficial moraines on early maps. Two fabric diagrams at the bottom of the figure are from microfabrics done on hand specimens. See Fig. 2 for location of sections and text for other localities.

measured 600 m from the west end of the island 4 m below the surface and contains a girdle along the eastern edge, with a slight maximum at about S 70° E. This may mean changes in ice flow direction from the east. Fabric 67-42-B GB, which was from 40 cm to 90 cm below the surface in a more clayey part of the till at the same section, suggests the same trend with a possible maximum in the opposite direction. This fabric is not strong and is useless in determining an individual flow direction. It has not been plotted on Fig. 10. This fabric may reflect an early flow from Casement Glacier to the northeast, modified by subsequent flow from the east. The other fabric from Adams island, 67-43 GB, is from the same unit as 67-42-B GB, 300 m to the west and about 1 m below the surface. This fabric has a strong easterly maximum which coincides with the direction indicated by the medial moraines (Fig. 3, 10). It also has a nearly transverse maximum at S 18° E with very little dip in this direction. Such transverse maxima have been interpreted as the result of interaction of pebbles within the glacier or attainment of the theoretical minimum energy state after a long time in the glacier (Glen and others, 1957); however, it is the writer's opinion that in this case it could also be the result of a change in flow direction of the ice. Observations and measurements made by the writer on pebbles in, and at the base of the ice, and on the ground in front of the Casement Glacier indicate a strong alignment of pebbles in all cases with the glacier flow direction. No indication of a transverse maximum was found.

Fabric 67-59 GB, from 50 cm to 90 cm below the surface in Glacier Bay till at Section 58, contained the highest concentration of pebbles. The till is about 110 cm thick over gravel at this point and the upper 50 cm are less indurated and more sandy than the lower part. The surface of the ground is covered with ablation till. The direction of flow indicated by the fabric is S 61° E toward Endicott Gap.

The fabric at Section 18 indicates a flow from N 40° E. This direction is almost parallel to the medial moraine about 600 m to the northwest (Fig. 3, 10). This fabric was measured at about 3 m depth in a till unit 4.2 m thick and, if the possibility of reorientation is omitted, it indicates an ice-flow during the early stages of deposition of the till similar to that in the latter stages of glaciation.

Two fabrics were measured at Section 69 on top of Adams mesa in Adams Valley. The till unit there is 2.7 m thick and contains a fine sand lens at about 1.5 m depth. Fabric 69-1 is from about 50 cm below the top and has a poorly defined maximum at about N 90° W; fabric 69-2 has a good maximum at N 40° W indicating ice flow into Adams Valley. The poor maximum of 69-1 is not surprising considering the reversal of flow that must have taken place. The final ice flow was out of the valley as indicated by medial moraines on Reid's map (Fig. 3), by the crag-and-tail on top of the mesa (Fig. 10), and by the lithologies of cobbles in ablation moraine derived from the upper part of Adams Glacier. The fabric direction, indicating glacier flow from Adams Inlet into Adams Valley, is supported by four pebble counts from Glacier Bay till units at this locality with an average of 25% plutonic igneous rocks. Although this rock type is confined to the area north of Adams Inlet these pebbles were probably derived in part from the underlying Berg Formation which now extends about 1 km south of Section 69. Other evidence of an advance from the north over this area are lake sediments on the Berg Formation and associated with several layers of till at the top of this section. This lake formed when drainage was blocked by ice from Adams Inlet. Two other facts support this hypothesis: (1) an analogy with the advance of ice in the Adams Formation; here also till fabrics indicate flow from the north as do the folded sediments beneath, and (2) the catchment basins of White and Adams Glaciers are low and small and probably retarded generation and advance of these glaciers compared to Muir and Casement Glaciers.

A microfabric (Fig. 27, Section 54-1) on till collected from the middle of a 2-m-thick unit in Section 54 indicates an ice-flow direction from the northeast out of Pyramid valley. A glacier at the head of the north tributary to the Pyramid river was probably the source of the material in this till.

Other Indications of Glacier Movements

Included in this part are brief descriptions of glacier movements as determined by striae, grooves, crag-and-tail, and roches moutonnées. The best developed striations and grooves in Adams Inlet occur on limestone; poor striations are found on hornfels and argillite. Ice-flow directions of striae and grooves were inferred from stoss and lee relationships, micro crag-and-tail, and relative roughness to the touch.

In the western part of the Inlet the flow direction indicated by striae parallels the directions of surficial moraines (Fig. 3, 10). Deviations (crossing striae) from this westward flow are believed to be due to late stage changes in ice-flow direction following thinning of Adams Inlet Glacier. This is particularly well developed near the peninsula in Adams Inlet (Fig. 10). Grooves, formed over a longer period of time than striae, in this peninsula are 2.4 m wide and 2 m deep. Indicated flow is also from the east.

East of Seal River the glacier flowed southwestward toward the Klotz Hills. Several of the more easterly striations (Fig. 10) indicate flow toward eastern Adams Inlet. Because of poor exposures and poor striations, the ice-flow directions north of Adams Inlet are not readily apparent. Westward flow out of Granite Canyon is indicated by striations and roches moutonnées at the entrance, but a southeasterly flow is also indicated. These different directions reflect the change of Granite Canyon Glacier from an active supplier of ice to Adams Inlet to a more passive condition when the Casement Glacier component forced Adams Inlet ice southward toward glacial Lake Endicott. Still later the ice-flow direction reversed and Adams Inlet ice probably flowed into Granite Canyon, which was already ice free in its central region.

At the east end of Adams Inlet near the entrance to Berg Valley, grooves and striations indicate a major flow direction toward Endicott Gap. Striations crossing this trend head into Berg Valley. At 250 m (820 ft) on a bedrock knob in the entrance to Berg Valley there is evidence of an earlier component of flow into this valley.

To the south of Goddess valley many limestone knobs with good stoss and lee surfaces and striae indicate flow toward Endicott Gap. Striations that cross these major ice-flow indicators were made during the latter stages of deglaciation as the ice retreated to the north and flow directions shifted. Some very faint and short striations that occur only on the highest part of bedrock knobs may have been formed by stranded icebergs in glacial Lake Endicott.

On the northwest corner of Tree Mountain at 420 m (1380 ft) several striations indicate flow around the mountain out of Adams Valley. The best example of a crag-and-tail feature occurs on top of Adams mesa. Here a 1.5-m-high boulder has a 30- to 40-cm-high tail of till on the lee (north) side. This feature was formed by ice moving north out of Adams Valley.

On the northeast corner of Tree Mountain striae occur at 810 m (2660 ft). This is below a scrub trim line and the striae were probably made during the last glaciation. The origin of striae on the head wall of the north cirque on Tree Mountain is not certain. They occur at 945 m (3100 ft) elevation 10 m below the top of the cirque. Another set, also on the horizontal surface of a 20 to 30 cm ledge, occurs 1 m below the top. These striae are about 150 m (500 ft) above the trim lines on either side of the mountain. It is possible that there was a divide in Adams Inlet Glacier at this position that raised the ice level in the cirque, but it probably did not fill it. The possibility of a cirque glacier is also unlikely because of the much lower trim lines near the edge of the cirque. Although the only tree in the trim line that was sampled was just over 100 years old, the trim line is so well developed that it is unlikely to have formed during a post-Neoglacial maximum stillstand at Endicott Gap. These striations were probably made by Wisconsin ice that filled the cirque and overrode Tree Mountain.

Trim Lines

In the map area trim lines, which are boundaries between vegetation (trees, shrubs, lichens) and rock, are best developed in Goddess valley from Adams Inlet to Endicott Gap. Here the main trim line varies from 760 m (2500 ft) on the east side of Tree Mountain and 700 m (2300 ft) on the east side of Goddess valley, to about 460 m (1500 ft) 1 km north of Endicott Lake. Surface slopes for the Endicott lobe range from 35 m per km (18 ft per mi) on the east side to 26 m per km (14 ft per mi) on the west side. This trim line was investigated on the east side of Tree Mountain where it is between 700 m (2300 ft) and 730 m (2400 ft)(Fig. 11). Above the trim line, mountain hemlocks (*Tsuga mertensiana*) are about

6 m high with diameters as much as 27 cm. One 14-cm-diameter tree that was cut 60 cm above the ground showed 103 annuli. About 10 m below the trim line a shrub-like mountain hemlock, 2.5 cm in diameter, contained 35 annuli. The trees above the trim line are probably those seen by Muir (1915) in 1880 and 1890, and Reid (1892) in 1890 and 1892. By 1890 the ice surface in Adams Inlet had already lowered to 380 m (1250 ft). If we use ice-surface-lowering rates of 6.1 m (20 ft) per year (Reid, noted in Field, 1947) and 7.9 m (26 ft) per year (Field, 1947), and assume the rates of lowering before and after 1890 to be about the same, then the Endicott lobe was still at the trim line in 1817 or 1842. It may have occupied this terminal position for some time prior to this. The climax of the Neoglacial advance of Bartlett Cove is estimated to have been about 1700 (Goldthwait and others, 1966).

Trim lines are also present on the mountains beside the lake of the headwaters of Goddess river. The best developed, and the one which probably corresponds to the main trim line in the Goddess valley, is between 520 m (1700 ft) and 550 (1800 ft). It represents the earlier, expanded position of Goddess glacier.

Other linear features of uncertain origin also occur on the map. The features mapped from air photographs may be poorly developed drainage channels, lateral moraines, or minor trim lines. They represent still-stands of ice margins. On the west side of Tree Mountain they are numerous and occur as low as 400 m (1300 ft) and as high as 760 m (2500 ft). Near Goddess glacier they occur up to 975 m (3200 ft), and may be features of the Wisconsin Glaciation.

Seal River Formation

Definition and Distribution

The Seal River Formation was defined by Goldthwait (Goldthwait and others, 1966) as the gravels of glacial outwash plains of the latest retreat of ice from the area. The formation was named after the Seal River which drains Casement Glacier to Adams Inlet. According to the expanded definition of the Glacier Bay Formation described earlier, pitted outwash and ice-contact gravels would not be included in the Seal River Formation. This formation is unit 7 of the standard succession given at the beginning of the chapter on glacial stratigraphy.

About 30 percent of the land surface below 300 m (1000 ft) is covered by gravels of these outwash channels. The largest deposits are those of the Seal, Goddess, and Adams Rivers (Plate I). Other smaller deposits of this formation are spread throughout the area wherever outwash has been, or is being, deposited.

Nature of Deposits

The gravels of the Seal River Formation are generally rounded, well-sorted, and uniformly bedded. They occur in outwash plains, terraces, deltas, and fans. These gravels were derived from material carried within glaciers, although in many cases this material has been incorporated from the underlying gravels (Upper Van Horn Formation to the north of Adams Inlet, and Berg Formation).

Thus some of the pebble roundness may be attributed to derivation from these water-transported deposits. Lithologically the Seal River Formation is most similar to the underlying Glacier Bay and Berg Formations in Goddess valley. There the PI/M ratios vary from only 0.31 to 0.35 (Table 2). The one sample of Seal River Formation from Adams Inlet has a ratio close to the average for the Glacier Bay till from Adams Inlet. The Seal River Formation in Granite Canyon and Seal Valley have PI/M ratios higher than other deposits in these areas. The average PI/M ratio for this formation in the Adams Inlet area is 2.3 (Table 2).

Post-Glacial Deposits

Alluvium

Included within this category of post-glacial deposits are flood-plain, fan, and lacustrine deposits. These deposits are of limited importance in the map area. They vary in mechanical composition from coarse gravel to fine silt. Alluvium is as much as 500 m wide in Granite Canyon and 300 m wide along Endicott River. The thickness here, as elsewhere in the Inlet, is not known. Other small patches of alluvium are associated with minor non-meltwater streams in the valleys. Some of these streams, such as those on the east side of Tree Mountain and the spring-fed streams in the "amphitheaters" on the east side of Goddess river, are building alluvial fans. Fan-deltas are common all along the south side of Adams Inlet.

Alluvium is also filling some of the lakes in the area. The unnamed headwater lake of Goddess river has become half-filled with alluvium in its island-dotted southeastern end since 1948. This represents an infilling of more than 500 m of gravel in the past 18 years.

Colluvium

Colluvial deposits in the area include talus, slides, mudflows, and rock stripes. The most widespread colluvial deposits are the talus deposits. They are most common in the area east of Casement Glacier, and on the north sides of Mt. Case and the unnamed mountain to the east. In this area south of the Inlet the talus blocks are 30 to 60 cm in diameter. The longest talus slope, on the north side of the unnamed mountain east of Mt. Case, is 2200 m long, and has an average slope of 26°. In places where measured stratigraphic sections (Fig. 2, sections 21, 61, and 79) are near talus slopes, talus material intermingles and in places dominates gravel deposits of the Berg Formation. Some of the talus is being deposited over a blanket of snow and/or a glacier, as is the case on both sides of the first southeast tributary valley of Casement Glacier. In this situation the upper part is talus; the lower part may more properly be termed an incipient lateral moraine.

Two slides were recognized in the area. On the west side of the first north-south U-shaped valley west of Granite Canyon (Plate I) is a 0.08 km^2 area that appears on the aerial photographs to be a rock slide. This slide is in an area of hornfelsed limy graywackes and shales. The slide overlies a fault (Fig. 5). The other slide area is on the north side of the terminus of Girdled Glacier. The debris from this slide, which occurs in brown marble and volcanics, covers an area of 0.23 km^2. A south-trending fault is 400 m west of the slide area (Fig. 5). The slide is also of the rockslide variety (Sharpe, 1938, p. 76), and apparently is still active. On July 12, 1967, at 8:30 a.m., a loud noise was heard at camp, and rising dust from Girdled Glacier signaled a new slide at this location. The slide area was not investigated after this event but had been seen more than a month earlier. At that time large boulders lay on the side and terminus of Girdled Glacier; boulders in the gorge immediately in front of the glacier were as much as 5 m in diameter.

Small mudflows occur in unconsolidated deposits on both sides of Goddess river. They are well developed near Section 41 (Fig. 28), where rhythmically bedded unstable silts and clays of the Adams Formation have flowed out from a steep 45-m-high slope into the fluvial terrace. The mudflows apparently remained stationary long enough to build up rims about 25 cm high and several meters in diameter. Later the original mudflow breached the rim and flowed out, leaving a low area behind the mudflow rim.

A debris flow was observed on the east side of Goddess river, between Sections 38 and 41 (Fig. 2). The flow, with a surface slope of 8° - 10°, consists of rounded and sub-rounded medium gravel. The upper 5 - 10 cm of the flow contains little interstratified material. Elongated in a northerly direction, the flow has dimensions of 19 m by 7 m. The front and sides are lobate with lobe heights of about 30 cm. A poorly sorted alluvial fan with approximately the same slope underlies the debris flow.

In the gully from which the flow apparently originated there is talus similar to that in the flow, but separated from it by 20 m of clayey and silty colluvium. The sides of the gully are topped by 6 m of gray medium-grained well-sorted gravel with rounded pebbles. This unit is underlain by 2.5 m of brown fine-grained and silty sand, which in turn is underlain by 12 m of gray rhythmically bedded clay and silt. This combination of deposits provided the material for the flow.

What apparently happened was a build-up of talus gravel over the colluvial clays and silts in the bottom of the gully. With a heavy rain, increased runoff through the gravels and over the clays finally resulted in an unstable mixture in the underlying and interbedded clays and silts (average mechanical composition of the Adams lacustrine units is 56.2 percent silt and 43.8 percent clay), and they flowed down the gully. Much mixing of the gravel and clay probably occurred; however, the upper 5 cm of the gravel flow now consists mainly of pebbles which may be the result of washing out of the finer particles.

A pebble fabric analysis was made on the upper surface of the flow. Three transects, each 3 m long, were made across the middle and all elongated pebbles, usually about 6-8 cm although one was 24 cm long, were measured. The fabric showed a maximum at N 62° W with a 9° upslope dip. The strike is off from the long axis of the flow, which is N 7° W. This difference is probably due to a diversion of material toward a lobe on the east side of the flow during the last stage of movement.

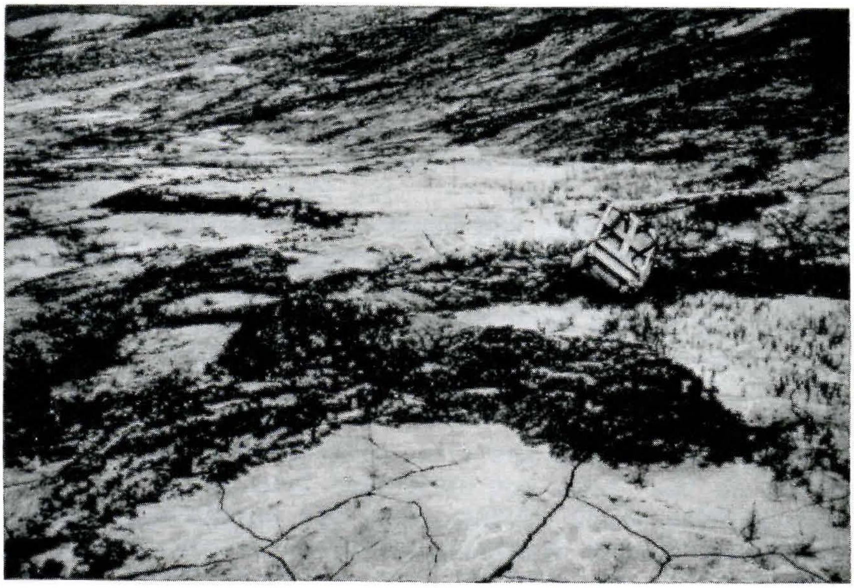


Figure 28. Mud-flow rims near Section 41 on the east side of Goddess river. Unstable silts and clays from gullies in the Adams Formation flowed into a fluvial terrace. The flow stabilized long enough to establish a forward rim but, with additional material the flow escaped from this dam to a lower level and repeated the process.

East of the top of Tree Mountain, at an elevation of about 1020 m (3350 ft), is a large area of unvegetated patterned ground. According to Washburn's (1956) classification of patterned ground this occurrence would be termed 'sorted stripes' because this pattern consists of lines of stones with finer material between. In this case the rows of coarse material (mostly less than 5 cm diameter but some fragments greater than 20 cm) are about 20 to 25 cm between centers. The maximum slope, which is approximately 10°, trends N 8° E. Large sorted stripes are known to occur in permafrost areas, but small ones may occur in permafrost-free ground (Washburn, 1956). A 30-cm-deep pit near the top of Tree Mountain did not contain permafrost; however, Ugolini found both stripes and permafrost near the top of Red Mountain (1091 m[3580 ft]), 20 km to the northwest (Goldthwait and others, 1966). At 900 m (1950 ft) on Mt. Wright, Ugolini also found permafrost 20 - 40 cm below the top of turf hummocks and 70 cm below the area between hummocks. Thus there is a good possibility that there is permafrost in the area of sorted stripes on Tree Mountain. Many mechanisms have been described for the origin of patterned ground, and Washburn (1956) favors a polygenetic origin for most features. These processes include desiccation, contraction due to low temperatures, local differential heaving, elevation and solifluction. More than one of these processes are probably active on Tree Mountain.

Eolian Deposits

Eolian deposits were noted at several localities in the map area. Many of these deposits consist of fine sand and occur adjacent to cliffs and unvegetated slopes of sand and gravel. Wind erosion of these cliffs lifts the sand and silt and spreads it over the uplands. On the surface of Adams mesa a sand deposit rests among boulders and cobbles of ablation moraine. Near Section 69 the eolian deposit is 3 cm thick. A gently sloping surface on the east side of Tree Mountain near Section 59 contains fine sand as thick as 2 cm blown from the face of the nearby cliff. Alder leaves here are sometimes covered with silt and fine sand. Near White River similar eolian deposits have been found and on several occasions dust clouds have been seen over the outwash of this valley. This loess is probably derived in large part from the high sand and gravel cliffs.

SEQUENCE OF EVENTS IN ADAMS INLET

Events Prior to Neoglacial Deglaciation

As is the case in the rest of Glacier Bay studied to date, Adams Inlet contains a record of late- and post-Wisconsin glacial history. During Wisconsin glaciation, ice overrode Tree Mountain (1028 m[3374 ft]) and probably reached higher elevations on Mt. Wright at the entrance to Adams Inlet. Many hills and mountains in the Endicott area were covered by ice. Glaciers filled Lynn Canal and Icy Strait to the south of Glacier Bay. During the Wisconsin stage, deposits of Granite Canyon Till and some deposits previously called Muir Formation, but now thought to be equivalent to the Granite Canyon Formation, were produced. Deglaciation of Adams Inlet was followed immediately by inundation and deposition of the Forest Creek glaciomarine clay and silt. This material is dated by peat from the top of the formation in Adams Inlet at 10,940 years B.P. (I-2395). At that time the land was depressed approximately 30 m relative to present sea level. Sometime during deposition of the Forest Creek Formation, a volcano, believed to be Mt. Edgecumbe 210 km south of Adams Inlet, erupted and spread ash over this area. Some of the ash was deposited directly in the sea; other quantities were washed from ash-covered slopes. A post-Wisconsin ice advance, possibly a pulsation of Casement Glacier, may have occurred north of Adams Inlet at Forest Creek where the Muir Formation overlies wood dated at 10,400 years B.P. (I-1615). There is a possibility that the Muir till is a solifluction or flow till deposit so this ice readvance is uncertain.

Events between deposition of the Forest Creek Formation and late stage (post-3850 B.P., I-3068) Lower Van Horn gravels are not well represented in the stratigraphic record of Adams Inlet, but may be inferred from the evidence in other parts of Glacier Bay. Emergence and outwash deposits followed deposition of the Forest Creek Formation so gravels overlie the Forest Creek Formation in Granite Canyon. Between 7050 B.P. and 4150 B.P. an intraglacial warm period, the Hypsithermal, produced mean annual temperatures in Wachusett and Muir Inlets about equal to those of the present (Goldthwait, 1966). Gravel probably was deposited in Adams Inlet at this time, although no dates are available in Adams Inlet for the period from about 10,000 to 4000 B.P. During early Neoglacial time 4150 to 2200 B.P. (Goldthwait, 1966), the Lower Van Horn gravels, derived from tributary valleys, continued to be deposited in Adams Inlet. Logs derived from trees on and adjacent to the Van Horn fluvial deposits in Adams Inlet were scattered along rivers in the tributary valleys. About 2000 B.P. at least one small short-lived lake had terminated and the subsequent forest, 20 m above present sea level, was overwhelmed by shifting outwash deposits by 1980 B.P. (I-2394). Glacial Lake Adams filled Adams Inlet and tributary valleys by 1700 B.P. (I-2687). Neoglacial ice advanced into this lake from north of Adams Inlet as much as 6 km in tributary valleys to the south and deposited till. Deposition of this lacustrine-till complex (Adams Formation) was followed by a retreat of ice that may be correlative with the Little Optimum (1150-1300 A.D.)(Lamb and others, 1966). A stillstand of Neoglacial ice in Adams Inlet resulted in deposition of outwash (Berg Formation) in the southern tributaries to an elevation of 250 m.

The late Neoglacial or Little Ice Age advance registered by the Glacier Bay till in Adams Inlet reached as far south as Endicott Gap where it remained until sometime between 1817 and 1842. Glaciers from two valleys south of the Inlet advanced into small lakes where they met the main Adams Inlet Glacier moving south into these valleys. Adams Inlet was then filled with ice to an elevation of about 760 m (2500 ft) and there was an ice divide between Muir Inlet and Endicott Valley.

Neoglacial Deglaciation

By 1890-92 when Reid (1896) produced the first detailed map (Fig. 3) of Adams Inlet, deglaciation had proceeded to the point where the ice was only 380 m (1250 ft) above tide in the Inlet. By this time lakes had formed in several tributary valleys. The largest proglacial lake was glacial Lake Endicott, which expanded northward from Endicott Gap in three stages, and finally drained into Adams Inlet by 1935.

Between 1890 and 1940 Adams Inlet Glacier ablated at a rate of 7.9 m per year determined by Field (1947). Cirques that were filled with ice in 1890 (Fig. 3) are ice-free today, or contain only stagnant remnants of glaciers as is the case in the cirque south of Tree Mountain (Fig. 29).

By 1929 Adams Inlet Glacier had receded from Endicott Gap to Berg Valley (Fig. 30). Glacial Lake Endicott was still dammed by this ice, and another lake probably existed in Granite Canyon. The higher portions of the peninsula in western Adams Inlet were exposed above the ice and the sea had invaded as far as the outwash of White Valley (Fig. 30). Adams Glacier was separated from Adams Inlet Glacier by a terminal moraine at the mouth of the valley. The changes in glacier margins, vegetation, and outwash that occurred between 1929 and 1967 are readily apparent on comparison of Fig. 30 and 31.

By 1941 Adams Inlet Glacier had an area of about 15 km^2 and its elevation was less than 100 m according to Field (1947). A small portion of the ice joined Adams island to the north side of the Inlet (Fig. 24). Some remnants of the surficial medial moraines on this ice mass are now distinguishable on the ground today. In 1941 Adams Glacier was still at the edge of Adams Inlet; by 1948 the glacier had receded 3 km to the south, and by 1967 it was 5 km from the Inlet. A considerable lowering of the surface of Adams Glacier (Fig. 29) accompanied this retreat.

While ice was stagnating in Adams Inlet and damming glacial meltwater and mountain streams to form lakes, it was also creating an environment favorable for the formation of eskers. Marginal meltwater streams

Figure 29. View south to Adams Glacier from near the top of Tree Mountain (1967). Cirque in valley to east was filled with ice to the dotted line in 1890 according to a photograph taken by H. F. Reid (1892, plate 5).

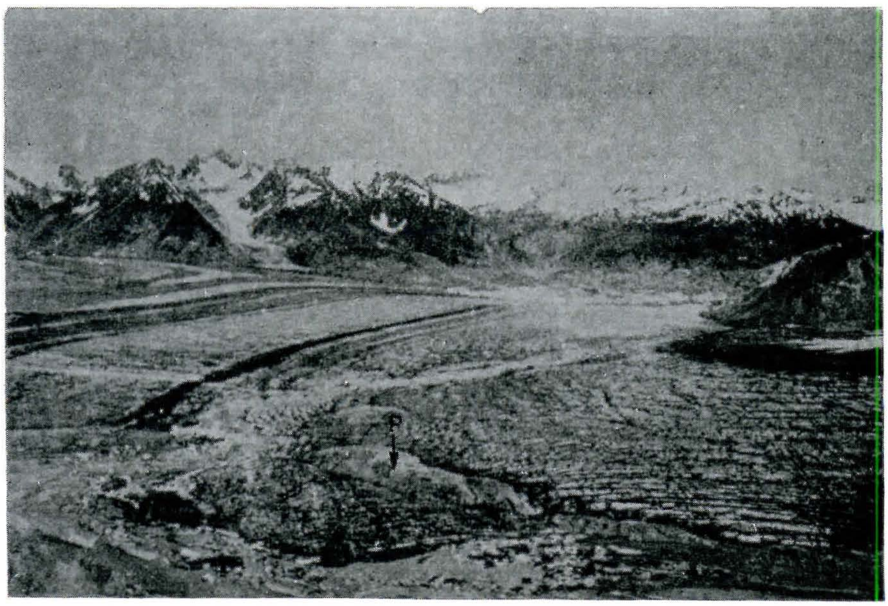


Figure 30. Adams Inlet in 1929. U. S. Navy photograph (Field, 1947) taken from over Klotz Hills. Peninsula (P) at the west end of Adams Inlet protrudes through the ice. Photograph courtesy of Geographical Review.

Figure 31. Adams Inlet in 1967 from north side of Mt. Case. Compare peninsula (P) with that in Fig. 30 above. Alluvial fan delta (D) on the south side of Adams Inlet extended 450 m since 1948. Girdled Glacier is in the center valley.

disappeared beneath the stagnating ice mass and deposited washed debris in the form of eskers. These are well developed along the north shore of Adams Inlet just below the ice margin of 1930. Shortly thereafter outwash from Seal, White, Granite, Goddess, and Adams Rivers began to accumulate in Adams Inlet. In some places where these rivers deposited outwash on or against wasting ice, terraces were cut upstream in response to lowered base level when the ice melted. In this same situation pitted outwash and kame terraces are now visible, with some buried ice still wasting away. Rivers carrying outwash and emptying into Adams Inlet have built several large outwash deltas, one of which has extended 300 m to 400 m in the past 19 years. Two outwash areas less than 1 km wide have extended 1 km and 1.3 km (50 to 70 m per year) over tidal flats in the same period.

Downstream from gullies in the thick unconsolidated deposits, and locally elsewhere, alluvial fans have been formed. Talus deposits have expanded to occupy valley walls bared by retreat of the glaciers. Wind transported sand and silt derived from bluffs of unconsolidated material, and to a lesser extent from outwash, have formed eolian deposits as thick as 3 cm on upland surfaces.

RELATIONSHIP OF DEPOSITS IN ADAMS INLET TO THOSE IN MUIR INLET

In order to compare the deposits in Muir and Adams Inlets composite stratigraphic sections for these two areas were reproduced (Fig. 32) with time lines, as far as they are known, included between the two sections. The section for Adams Inlet is a modified version of Fig. 6; the section for Muir Inlet is taken from Haselton (1966). A brief discussion of some of the differences between the stratigraphic sections is given below.

The Adams Inlet glacial section is much thicker than that in Muir Inlet, despite the fact that it represents a relatively short period of time. Although there are dated deposits as old as 11,170 ± 225 years B.P. (I-2396) in Adams Inlet, most of the stratigraphy exposed in the Inlet is less than 3700 years old. In Muir Inlet, deposits as old as 7000 years commonly occur, but the oldest deposits are dated at only 10,400 ± 260 years B.P. (I-1615).

Comparison of the stratigraphic columns (Fig. 32) of the two areas reveals that the difference in the succession is mainly between the Van Horn Formation in Muir Inlet and the Adams and Berg Formations in Adams Inlet. The Adams Formation, consisting of as much as 75 m of lacustrine material and till, is younger than 1700 years B.P. It represents the deposits of a lake into which ice advanced several times from north of Adams Inlet. This advance is apparently not recorded in other parts of Glacier Bay and indeed ice already lay continuously over upper Muir and Wachusett Inlets as this occurred. While this event was taking place, outwash gravel and till of the Van Horn and Glacier Bay Formations, respectively, continued to be deposited in several parts of Muir Inlet. The establishment of a new formation for the lacustrine-till complex, despite some similarity between it and others, is necessary because the nature of the deposits is significantly different from either the middle member of the Van Horn Formation or the Glacier Bay till.

The Berg Formation is about 90 m thick at the maximum and consists of medium- to coarse-grained sand and deltaic gravels. It differs from the upper member of the Van Horn Formation in that it contains more sand and the gravels are mainly deltaic deposits. The Berg Formation is the outwash from ice that retreated from Lake Adams and later readvanced over Adams Inlet. The Berg Formation was probably formed several hundred years after the deposition of the upper member of the Van Horn Formation, even in the lower part of Muir Inlet where the Van Horn Formation is youngest.

The lower member of the Van Horn Formation in Muir Inlet is considered to extend into Adams Inlet. It represents the infilling of valleys by outwash from glaciers and by fluvial materials eroded from the mountains and older glacial deposits in the area. This reworked mode of origin is particularly true of much of the material derived in Adams Inlet from the south. In Muir Inlet the lower member of the Van Horn gravel is identified by its dark brown color, the lithology being highly

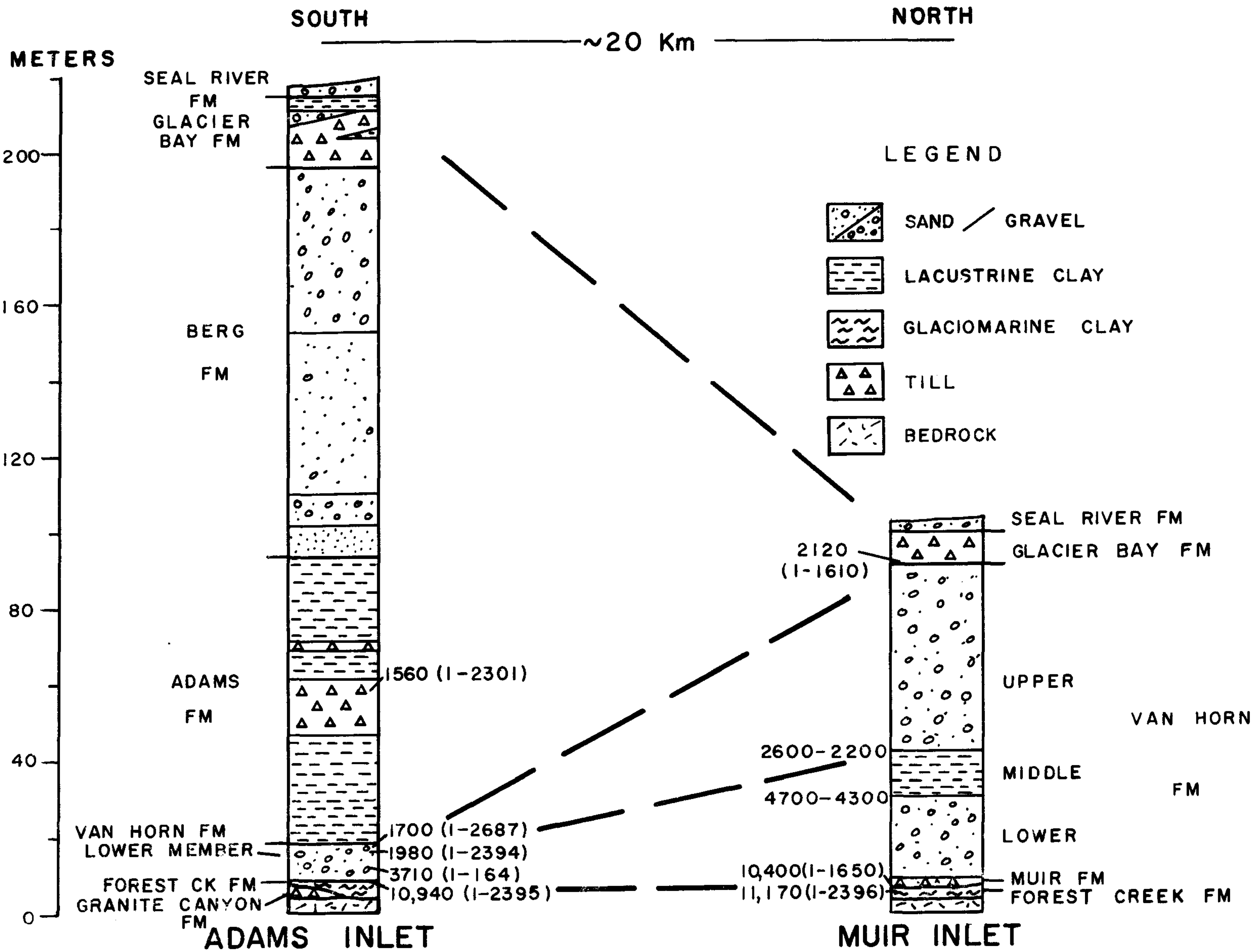


Figure 32. Comparison of stratigraphy in Adams and Muir Inlets; the section for Muir Inlet is from Haselton (1966). Dashed lines join deposits of approximately equivalent time. All radiocarbon dates are years B.P.; errors in the dates are given in Table 6.

variable depending on the tributary valley source. In Muir Inlet the age of the lower member of the Van Horn ranges from about 7000 B.P. to 4700 - 4300 B.P.; in Adams Inlet the lower Van Horn gravels are greater than 1700 years B.P. with the oldest date being 3710 ± 110 years B.P. (I-164) based on a sample of wood taken from about 3 m above tide on the south shore of the Inlet. Some gravels of the formation in Adams Inlet are much older. The gravels in Adams Inlet are, like those in Muir Inlet, varied in lithology and have a characteristic brown color. In one part of Adams Inlet, Granite Canyon, there are weathered light brown gravels overlying the Forest Creek Formation. The date at the interface is 11,170 years B.P. These gravels are also considered to belong to the lower member of the Van Horn Formation.

Forest Creek Formation deposits are probably very nearly equivalent in age in Adams and Muir Inlets. The presence of the Granite Canyon Formation in Adams Inlet only and the Muir Formation in Muir Inlet only constitute the main differences between the lower stratigraphic columns of these areas. This study shows that the Granite Canyon Formation is present at the base of the sedimentary succession in Muir Inlet. Glacio-lacustrine and glaciofluvial material included within the expanded definition of the Glacier Bay Formation constitute the main difference in the upper part of the stratigraphic columns of the two areas.

SUMMARY

Adams Inlet, in the eastern part of Glacier Bay National Monument, is in the rolling and rugged Chilkat-Baranof Mountains that reach heights of 1680 m (5510 ft) in the vicinity of the Inlet. In 1967, about 14 percent of the Juneau D-6 Quadrangle, the area of study, consisted of glaciers, and 30 percent was unconsolidated deposits in the lowlands. The climate of the area is a cool snow-forest climate, moist in all seasons, with cool summers (Köppen's 'Dfc' climate). Paleozoic clastic and carbonate sediments, metasediments, volcanics, metavolcanics and intrusions comprise the rock types in the vicinity of the Inlet. Plutonic igneous rocks are restricted to the area north of Adams Inlet.

Adams Inlet holds the key to the earliest and some of the latest events in the glacial history of the Monument; details of the Hypsithermal (7000 to 4000 B.P.) are not dated in deposits of the Adams Inlet area. The principal surficial deposits of the area are, from oldest to youngest: Granite Canyon Till and Forest Creek Formation, prior to 10,000 years ago, and lower member of the Van Horn Formation, Adams Formation, Berg Formation, Glacier Bay Formation, and Seal River Formation, from 4000 B.P. to the present.

The Granite Canyon till was deposited in Adams Inlet by Wisconsin ice. Deglaciation was followed by marine invasion to elevations 30 m or more above present sea level. The fossil assemblage of the glaciomarine Forest Creek Formation indicates water depths of 2 to 20 m. By 10,940 years B.P. uplift had proceeded to the point where ferns, sedges, and grasses became established on the marine sediments. A volcanic ash layer in the Forest Creek Formation is thought to have been derived from an eruption of Mt. Edgecumbe on the basis of age and proximity to a similar deposit near Juneau that is tentatively correlated with Mt. Edgecumbe.

The status of the Muir Formation is in question. At Forest Creek where the Muir Formation overlies wood (dated at 10,400 years B.P., I-1615) and the Forest Creek Formation, the till may be a solifluction deposit, flow till, or local till from a pulsation of nearby Casement Glacier. It is not thought now to represent a major post-Wisconsin advance in Glacier Bay. Other deposits of the Muir Formation are believed to be equivalent in age to the Granite Canyon till.

Marine sedimentation was followed by a period of gravel filling of the valleys around Adams Inlet. About 3700 B.P. gravels from tributary valleys had reached an elevation of 3 m in the area of Adams Inlet. Short-lived spruce forests, with some trees 60 cm in diameter, occupied the flood plains of these tributary valleys. Further north these were clearly on outwash, but here may have been remote from glaciers. In Muir Inlet between about 4500 and 2200 B.P. lacustrine sediments were deposited in several lakes, and in Wachusett Inlet there was a regular buildup of outwash. To the north of Adams Inlet advancing glaciers

reached Wachusett Inlet by 2735 B.P. (I-122) and topped White Thunder Ridge in Upper Muir Inlet by 2120 B.P. (I-1610). Advancing ice in Glacier Bay apparently reached Reid Inlet by about 4680 B.P. (Y-9). By 1700 B.P. (I-2687) outwash from glaciers in Muir Inlet and/or ice from Glacier Bay (Russell drainage) dammed the entrance to Adams Inlet forming glacial Lake Adams. The ice-dam hypothesis is supported by a date of 1540 B.P. (Y-4) on wood under till on the south side of Geikie Inlet. During this lake stage several advances of Neoglacial ice from north or west of Adams Inlet moved into the tributary valleys to the south and deposited the Adams till. A retreat of this ice less than 1100 years ago to the central area of Adams Inlet was followed by deposition of large quantities of gravel and sand (Berg Formation) in the valleys south of the Inlet. This stillstand is not dated radiometrically but may correlate with the Little Optimum (1150-1300 A.D.). Late Neoglacial ice advanced as far south in Endicott valley as Endicott Lake, where it remained until about 1830.

By 1890 when the first detailed map of Adams Inlet was made the ice was at an elevation of 380 m (1250 ft), about half of the elevation of the Neoglacial maximum. Several ice-dammed lakes had formed in tributary valleys. By about 1935, the largest of these lakes, glacial Lake Endicott, had drained into the sea. The rate of disappearance of the ice from 1890 to 1940 over the island in Adams Inlet has been determined by Field (1947) to be 7.9 m per year. Lateral channels, lateral moraines, and trim lines mark the former positions of the ice margins.

The Glacier Bay Formation is here expanded to include both the unsorted and sorted ice-contact deposits. Sorted deposits include kames, eskers, crevasse fillings, and pitted and collapsed outwash, which are common in this area of stagnating ice.

Methods employed in the differentiation and correlation of various units within Adams Inlet include: pebble counts, grain-size analyses, carbonate analyses, clay mineral determinations, and partial elemental analyses. In the pebble counts the plutonic igneous to metasediment plus sediment (PI/M) ratios were found to be helpful in differentiation of the tills, and in determining the source areas of tills and other units. The differences in the average PI/M ratios for the Granite Canyon, Adams and Glacier Bay tills is significant for the immediate area of Adams Inlet; however, when the tributary valleys are included in the average the difference between the Adams and Glacier Bay tills is not significant according to Student's $\underline{t}$ test. Large differences in the pebble lithologies in one unit may occur from valley to valley.

Differentiation of the three tills of the area was possible using grain-size analyses of the -2 mm fraction. This method is thought to be the most reliable for differentiation of the tills. Several elements determined in the X-ray spectrographic analysis of the sand and silt fraction also suggest differences in the till units.

Only limited success was obtained by using clay minerals and carbonates to differentiate the tills. This is in part due to the limited number of samples used in these determinations. Clay minerals occurring in most of the units include illite, chlorite, montmorillonite, and interstratified clays. Total carbonate occurring in tills in Adams Inlet varied from 7.8 to 13.6 percent. More determinations of the carbonate content of the tills might show significant differences between the units.

In correlating the Glacier Bay till in Adams Inlet with that in Muir Inlet, surprisingly good results were obtained with pebble counts and grain-size analyses, the two methods for which data were available for Muir Inlet. For samples from Muir Inlet the PI/M ratio is 0.67; for samples from Adams Inlet the ratio is 0.70. The average mechanical compositions are almost the same for the two areas. These data indicate that the source of most of the Glacier Bay till in Adams Inlet is from north of the Inlet.

Future investigations into the glacial history of Glacier Bay National Monument should be directed toward the problem of the positions of glaciers in Glacier Bay about 4000 and 1500 years ago. Knowledge of glacier termini at these times would help to explain the development of the lakes in Adams and Muir Inlets.

A detailed pedological examination and pebble fabric study of the Muir till at Forest Creek are needed to determine the exact nature of this deposit. In conjunction with such a study a search should be made for deposits indicating an advance of glaciers after 10,400 years B.P., or evidence for continuous ice cover from the Wisconsin to 8000 or 9000 years ago in other parts of Glacier Bay National Monument. Such information would be useful in the correlation of worldwide fluctuations of glaciers.

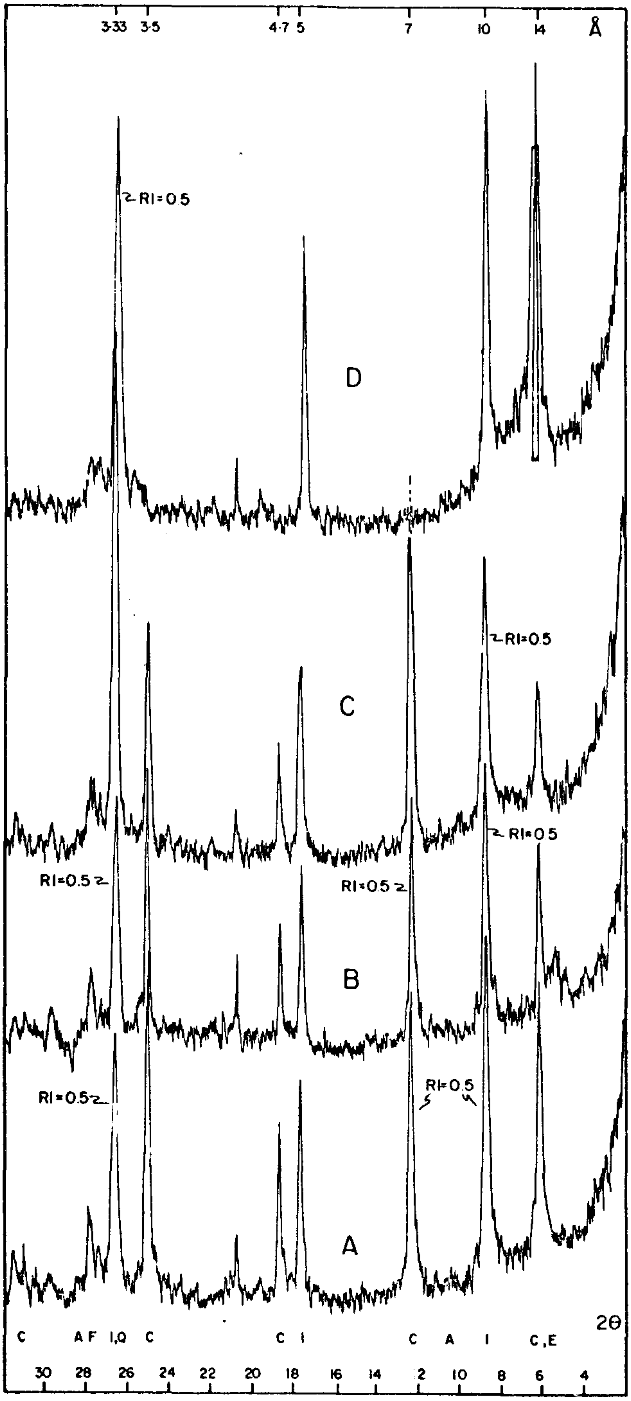
A study of the clay minerals now being deposited in Adams Inlet may be useful in determining the environments of deposition of the glaciolacustrine and glaciomarine deposits such as the Adams and Forest Creek Formations, respectively. Glacier Bay would also be a good place to pursue a study on the environments of deposition near tidewater glaciers. A detailed study, using submersible vehicles, of the deposits now forming beneath and in front of floating glaciers could be related to a coordinated study of glaciomarine deposits now elevated above sea level in Glacier Bay.

A continuation of the meteorological studies in Glacier Bay and particularly in Adams Inlet would aid in understanding the climate of this recently deglaciated area. Such information would be useful for studies of reforestation of the newly exposed land and also for studies of ablation of buried ice in Adams Inlet. Perhaps the studies most critical to the reasons for and duration of glacier retreat will be winter snowfall, summer melt, and net accumulation in the high surrounding icefields as already begun by Peterson (1969).

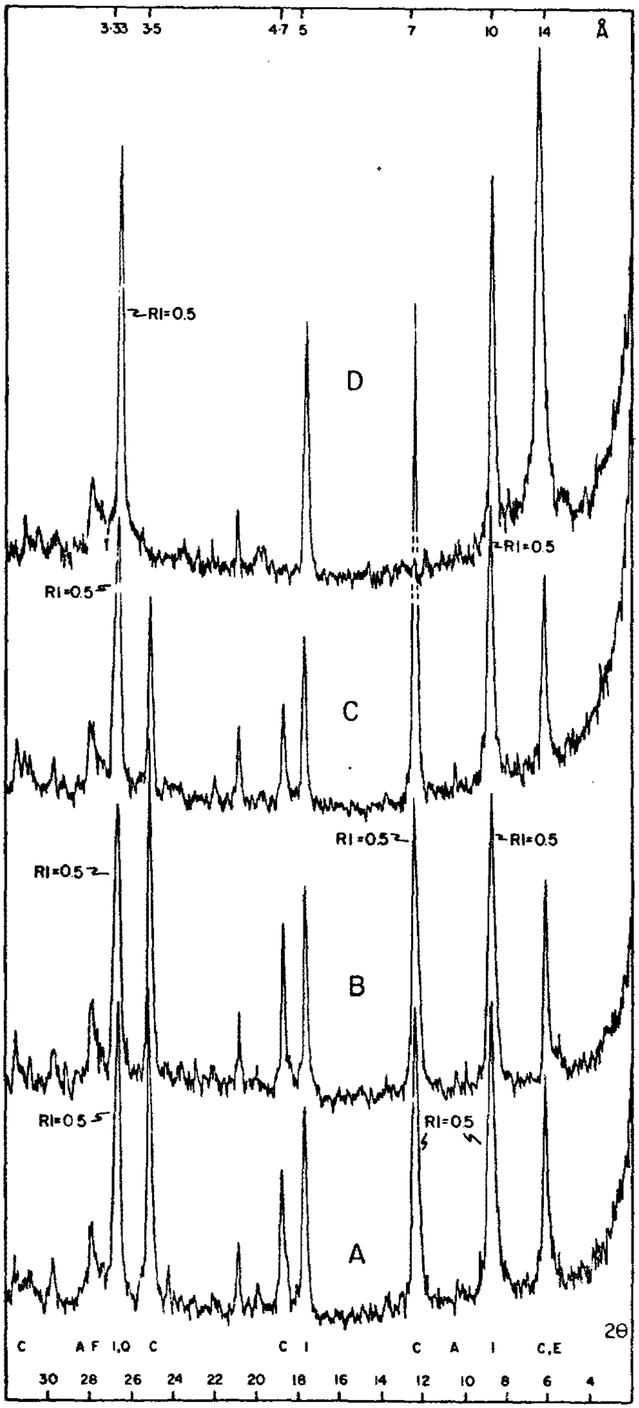
APPENDIX

Selected X-ray diffractograms of clays from Adams Inlet. Samples include: weathered and unweathered Granite Canyon till, Forest Creek glaciomarine clay, and Adams lacustrine clay.

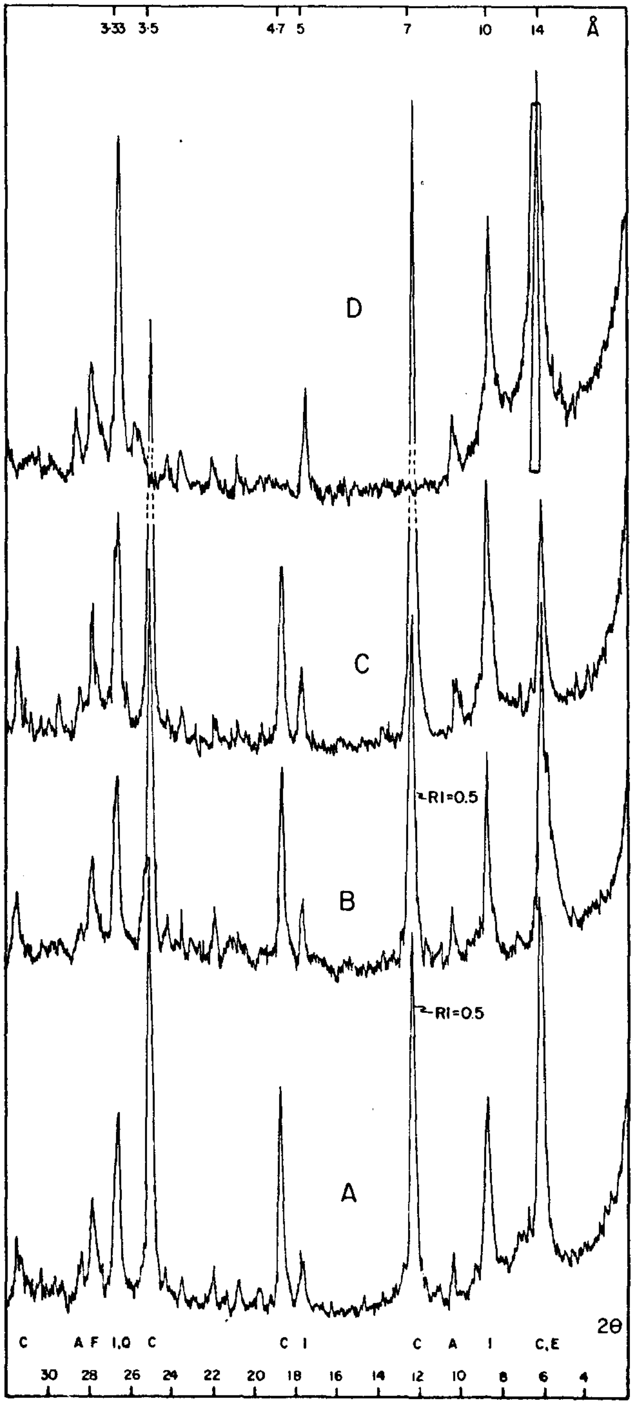
All tracings are of the -2μ fraction of the material; all samples are homoionic in magnesium. Letters identify the treatment: A - air dry, B - ethylene glycol, C - 400°C, D - 550°C. Degrees 2θ are $CuK\alpha$ radiation. R.I. is relative intensity. Symbols identifying the peaks: C - chlorite, I - illite, E - expandable clay, A - amphibole, F - feldspar, Q - quartz.



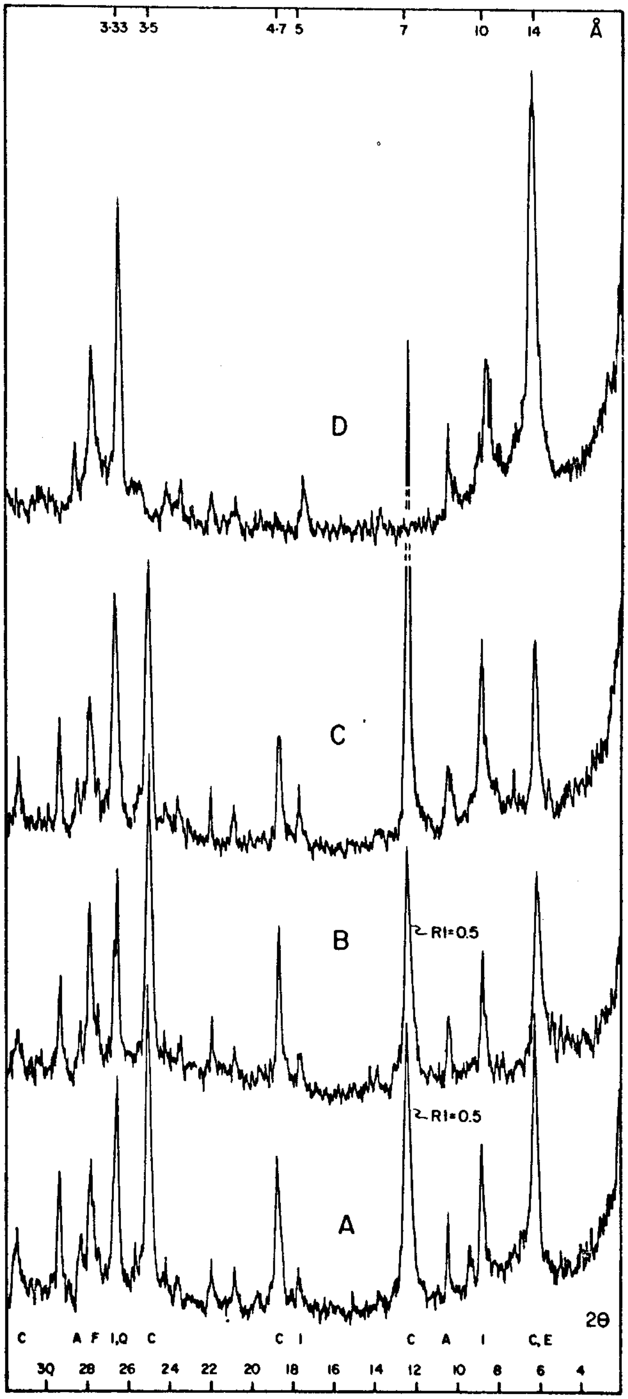
1. X-ray diffractograms of sample 66-8, weathered Granite Canyon till.



2. X-ray diffractograms of sample 66-9, unweathered Granite Canyon till.



3. X-ray diffractograms of sample 67D-8, Forest Creek glaciomarine clay.



4. X-ray diffractograms of sample 58-8, Adams lacustrine clay.

REFERENCES

Alaskan Boundary Tribunal, 1904, Atlas of Award: U.S. Gov't Printing Office, Washington, D.C., Map sheet 14.

Beavers, A. H., 1960, Use of X-ray spectrographic analysis for the study of soil genesis: 7th Intern. Congress of Soil Science, v. 2, p. 1-9.

Biscaye, P. E., 1965, Mineralogy and sedimentation of recent deep-sea clay in the Atlantic Ocean and adjacent seas and oceans: Geol. Soc. America Bull., v. 76, p. 803-832.

Carroll, D., and Starkey, H. C., 1959, Effect of sea-water on clay minerals, in Swineford, Ada, ed., Clays and Clay Minerals: Pergamon Press, New York, p. 80-101.

Colinvaux, P. A., 1967, Quaternary vegetational history of Arctic Alaska, in Hopkins, D. M., ed., The Bering Land Bridge: Stanford Univ. Press, Stanford, California, p. 207-231.

Cooper, W. S., 1931, A third expedition to Glacier Bay, Alaska: Ecology, v. 12, p. 61-95.

_______________ 1937, The problem of Glacier Bay, Alaska: A study of glacier variations: Geogr. Rev., v. 27, p. 37-62.

Cushing, H. P., 1891, Notes on the Muir Glacier region: Amer. Geologist, v. 8, p. 207-230.

Dreimanis, A., and Reavely, G. H., 1953, Differentiation of the lower and upper till along the north shore of Lake Erie: Jour. Sed. Petrology, v. 23, p. 238-259.

Field, W. O. Jr., 1947, Glacier recession in Muir Inlet, Glacier Bay, Alaska: Geogr. Rev., v. 37, p. 369-399.

_______________ 1964, Observations of glacier variations in Glacier Bay, southeastern Alaska, 1958 and 1961: Preliminary Report, Amer. Geogr. Soc., 35 p.

Flint, R. F., Sanders, J. E., and Rodgers, John, 1960, Diamictite, a substitute term for symmictite: Geol. Soc. America Bull., v. 71, p. 1809-1810.

Folk, R. L., 1961, Petrology of sedimentary rocks: Hemphill's Book Store, Austin, Texas, 154 p.

George, W. O., 1924, The relation of the physical properties of natural glasses to their chemical composition: Jour. Geol., v. 32, p. 353-372.

Glen, J. W., Donner, J. J., and West, R. G., 1957, On the mechanism by which stones in till become oriented: Amer. Jour. Sci., v. 255, p. 194-205.

Goldthwait, R. P., 1963, Dating the Little Ice Age in Glacier Bay, Alaska: Intern. Geol. Cong. 21, 1960, pt. 27, p. 37-46.

______________ 1966, Evidence from Alaskan glaciers of major climatic changes, in Sawyer, J. S., ed., Proc. Intern. Symp. on World Climate from 8000 to 0 B.C.: Royal Met. Soc., London, p. 40-53.

Goldthwait, R. P., Loewe, F., Ugolini, F. C., Decker, H. F., DeLong, D. M., Trautman, M. B., Good, E. E., Merrell, T. R. III, and Rudolph, E. D., 1966, Soil development and ecological succession in a deglaciated area of Muir Inlet, southeast Alaska: Ohio State Univ. Inst. Polar Studies, Rept. 20, 167 p.

Griffen, G. M., 1962, Regional clay mineral facies--products of weathering intensity and current distribution in the northeastern Gulf of Mexico: Geol. Soc. America Bull., v. 73, p. 737-768.

Griffen, J. J., and Goldberg, E. D., 1963, Clay-mineral distribution in the Pacific Ocean, in Hill, M. N., ed., The sea, v. 3 Intersci. Pub., New York, p. 728-741.

Griffiths, J. C., 1967, Scientific method in analysis of sediments: McGraw Hill, New York, 508 p.

Grim, R. E., and Johns, W. D., 1954, Clay mineral investigation of sediments in the northern Gulf of Mexico, in Swineford, Ada, and Plummer, N. V., eds., Clays and clay minerals: Natl. Res. Counc. Pub. 327, p. 81-103.

Harrison, P. W., 1957, A clay till fabric: its character and origin: Jour. Geol., v. 65, p. 275-308.

Haselton, G. M., 1966, Glacial geology of Muir Inlet, southeast Alaska: Ohio State Univ. Inst. Polar Studies, Rept. 18, 34 p.

______________ 1967, Glacial Geology of Muir Inlet, southeastern Alaska: Ph.D. Dissertation, The Ohio State University, 228 p.

Heusser, C. J., 1960, Late-Pleistocene environments of North Pacific North America: Amer. Geogr. Soc., Sp. Pub. 35, 308.

Hicks, S. D., and Shofnos, W., 1965, The determination of land emergence from sea level observations in southeastern Alaska: Jour. Geophys. Res., v. 70, p. 3315-3320.

Inman, D. L., 1952, Measures for describing the size distribution of sediments: Jour. Sed. Petrology, v. 22, p. 125-145.

Jackson, M. L., Tyler, S. A., Willie, A. L., Bourbeau, G. A., and Pennington, R. P., 1948, Weathering sequence of clay-size minerals in soils and sediments: I Fundamental generalizations: Jour. Phys. Coll. Chem., v. 52, p. 1237-1260.

Kamb, W. B., 1959, Ice petrofabric observations from Blue Glacier, Washington in relation to theory and experiment: Jour. Geophys. Res., v. 64, p. 1891-1909.

Keller, W., 1963, Diagenesis in clay minerals--a review, in Bradley, W. F., ed., Clays and clay minerals, Pergamon Press, New York, p. 136-157.

Lamb, H. H., Lewis, R. P. W., and Woodroffe, A., 1966, Atmospheric circulation and the main climatic variables between 8000 and 0 B.C.: meteorological evidence, in Sawyer, J. S., ed., Proc. Intern. Symp. on World Climate from 8000 to 0 B.C.: Royal Met. Soc. London, p. 174-217.

Lawrence, D. B., 1958, Glaciers and vegetation in southeastern Alaska: Amer. Scientist, v. 46, p. 89-122.

Lindsay, John, 1966, Observations on the level of a self-draining lake on the Casement Glacier, Alaska: Jour. Glaciology, v. 6, p. 443-445.

MacClintock, P., and Dreimanis, A., 1964, Reorientation of till fabric by overriding glacier in the St. Lawrence Valley: Amer. Jour. Sci., v. 262, p. 133-142.

Mackenzie, F. T., and Garrels, R. M., 1966, Chemical mass balance between rivers and oceans: Amer. Jour. Sci., v. 264, p. 507-525.

MacKevett, E. M., Jr., Brew, D. A., Hawley, C. C., Huff, L. C., and Smith, J. G., 1967, Mineral resources of Glacier Bay National Monument, Alaska: Open-file report, U.S. Geol. Survey, 176 p.

Mannerfelt, C. M.: son, 1949, Marginal drainage channels as indicators of the gradients of Quaternary ice caps: Geogr. Ann., v. 31, p. 194-199.

McKenzie, G. D., 1968, Glacial history of Adams Inlet, southeastern Alaska: Ph.D. Dissertation, The Ohio State University, 200 p.

Merrell, T. R., III, in preparation, Wildlife in Adams Inlet and vicinity, The Ohio State Univ. Inst. Polar Studies.

Muir, J., 1915, Travels in Alaska: Houghton-Mifflin Co., Boston, 329 p.

Orth, D. J., 1967, Dictionary of Alaska place names: U.S. Geol. Survey Prof. Paper 567, 1084 p.

Ovenshine, A. T., 1967, Provenance of recent glacial ice in lower Glacier Bay, southeastern Alaska: U.S. Geol. Survey Prof. Paper 575-D, p. 198-202.

Peterson, D. N., 1968, Observations on basal sliding of the Casement Glacier, southeast Alaska: (abs) Amer. Geophys. Union Trans., v. 49, p. 167.

_____________ 1969, Glaciological investigations on the Casement Glacier, southeast Alaska: Ph.D. Dissertation, The Ohio State University, 183 p.

Porter, S. C., 1966, Pleistocene geology of Anaktuvuk Pass, Central Brooks Range, Alaska: Arctic Inst. North America Tech. Paper No. 18, 100 p.

Porter, S. C., and Denton, G. H., 1967, Chronology of Neoglaciation in the North American Cordillera: Amer. Jour. Sci., v. 265, p. 177-210.

Powers, M. C., 1957, Adjustment of land-derived clays to the marine environment: Jour. Sed. Petrology, v. 27, p. 355-372.

Price, R. J., 1964, Land forms produced by the wastage of the Casement Glacier, southeast Alaska: Ohio State Univ. Inst. Polar Studies, Rept. 9, 24 p.

_____________ 1965, The changing proglacial environment of the Casement Glacier, Glacier Bay, Alaska: Trans. and Papers, Inst. British Geographers, Publ., no. 36, p. 107-116.

_____________ 1966, Eskers near the Casement Glacier, Alaska: Geogr. Ann., v. 48, Ser. A., p. 111-125.

Reed, J. C., 1938, Some mineral deposits of Glacier Bay and vicinity, Alaska: Econ. Geol., v. 33, p. 52-80.

Reid, H. F., 1892, Studies of Muir Glacier, Alaska: National Geogr. Mag., v. 4, p. 19-84.

_____________ 1896, Glacier Bay and its glaciers: U.S. Geol. Survey Annual Rept. 16, 1894-1895, pt. I, p. 415-461.

Rossman, D. L., 1963, Geology of the eastern part of the Mount Fairweather Quadrangle, Alaska: U.S. Geol. Survey Bull. 1121-K, 55 p.

Sharpe, C. F. S., 1938, Landslides and related phenomena: Columbia Univ. Press, N. Y., 137 p.

Shepard, F. P., 1963, Thirty-five thousand years of sea level, in Clements, T., ed., Essays in marine geology: Univ. S. Calif. Press, Los Angeles, p. 1-10.

Smeck, N. E., Wilding, L. P., and Holowaychuk, N., 1968, Genesis of argillic horizons in Celina and Morley soils of western Ohio: Soil Sci. Soc. America Proc., v. 32, p. 550-556.

Taylor, L. D., 1962, Ice structures, Burroughs Glacier, southeast Alaska: Ohio State Univ. Inst. Polar Studies, Rept. 3, 106 p.

Thorp, James, Cady, J. G., and Gamble, E. E., 1959, Genesis of Miami silt loam: Soil Sci. Soc. America Proc., v. 23, p. 65-70.

Todd, Ruth, and Low, Doris, 1967, Recent Foraminifera from the Gulf of Alaska and southeast Alaska: U.S. Geol. Survey, Prof. Paper 573-A, 46 p.

Twenhofel, W. S., 1946, Molybdenite deposits of the Nunatak area, Muir Inlet, Glacier Bay: U.S. Geol. Survey Bull. 947-B, p. 9-18.

__________ 1952, Recent shoreline changes along the Pacific coast of Alaska: Amer. Jour. Sci., v. 250, p. 523-548.

Vancouver, G., 1801, A voyage of discovery: London, v. 5, p. 421-422.

Wahrhaftig, C., 1965, Physiographic divisions of Alaska: U.S. Geol. Survey, Prof. Paper 482, 52 p.

Washburn, A. L., 1956, Classification of patterned ground and review of suggested origins: Geol. Soc. America Bull., v. 67, p. 823-866.

Weaver, C. E., 1959, The clay petrology of sediments, in Swineford, Ada, ed., Clays and clay minerals: Pergamon Press, New York, p. 154-187.

Welch, R. A., 1964, The form and origin of landforms produced during the wastage of Casement Glacier, Alaska: Master's Thesis, Univ. of Oklahoma, 95 p.

__________ 1965, Ecological observations in the Muir Inlet area, Glacier Bay National Monument, Alaska: Ohio State Univ. Inst. Polar Studies, Rept. 15, 19 p.

Wright, F. E., and Wright, C. W., 1937, The Glacier Bay National Monument in southeastern Alaska, its glaciers and geology: Open-file report, U.S. Geol. Survey.

Wright, G. F., 1887, The Muir Glacier: Amer. Jour. Sci., v. 33, p. 1-18.

__________ 1891, The ice age in North America: D. Appleton and Co., New York, 648 p.

Wright, J. R., Leahy, A., and Rice, H. M., 1959, Chemical, morphological and mineralogical characteristics of a chronosequence of soils on alluvial deposits in the Northwest Territories: Canadian Jour. Soil Sci., v. 39, p. 32-43.